

PETROLOGY OF THE IGNEOUS ROCKS

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PREFACE

THE first edition of this book, a slender volume of 128 pages, written by the late Dr. F. H. Hatch, O.B.E., was published in 1891. In 1926 one of us (A. K. W.) joined Dr. Hatch in a complete revision. Ten years later, after the death of Dr. Hatch, more drastic alterations were made in the scheme of classification of the igneous rocks, mode of occurrence being dropped in favour of grain-size—degree of crystallinity, while in the tenth edition another of the main props of former classifications—silica percentage—was discarded in favour of the mineral contents of the rocks. The reasons for these fundamental changes are discussed in the text; but as a consequence of them, and partly due to the growth of knowledge, the subject-matter of the original book has now been completely replaced, and the illustrations are entirely new. Although the authors feel justified in claiming that the work is now entirely their own, it is considered expedient to retain the sub-title "Hatch and Wells," which has become familiar through usage by several generations of students.

The chief alterations in the new edition have resulted largely from recent important advances in experimental aspects of petrology (using the term in the widest sense), notably in X-ray crystal chemistry, in equilibrium studies of silicate melts involving water as a component, and in the systematic analysis of individual rocks and rock suites. Some experimental results are of such fundamental significance that they merit inclusion in even an elementary discussion of petrogenesis: the determination of the temperature of formation of granitic melts involving water as a component; data relating to the process of serpentinization; and the discovery that the components of calcite in the presence of adequate water form low-temperature melts, are cases in point.

To an increasing degree it is becoming recognized that in petrology, laboratory studies and book knowledge must be supplemented by detailed studies of field- and structural-relationships. This tendency has led us to consider in more detail than formerly the structures of igneous rocks, and in a later part of the book, the account of igneous phenomena in Britain throughout geological time.

The authors' teaching experience has convinced them that, although theories concerning the modes of origin are particularly attractive to many students, the fundamental need is knowledge of the rocks themselves. Such knowledge provides the only firm foundation upon which theories of origin may be built. In this book, therefore, emphasis is placed upon petrographic aspects of petrology: features of the mineral composition and textures of the rocks which the reader can verify by his own observation.

In former editions certain aspects of petrogenesis received some

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attention; but in the present one, instead of relegating these matters to a special chapter towards the end of the book they are introduced at appropriate points in the text. Thus the petrographic description of each important rock-group includes, or is followed by, an account of the distribution of the rock-types concerned and a discussion of their origin. In this respect the authors believe the book to be indeed Petrology rather than Petrography.

Special attention has been devoted to nomenclature and definitions of rock-types with the object of eliminating as many unnecessary and uninformative names as possible. Unfortunately the number continues to grow, and in the last few years angolaite, tuvinit, busonite, uncompahgrite, niligongite, ankaramite as well as ankaramite (surely specially devised for the mystification of students) and even bugite have been added to the list already running into several hundreds. It is highly desirable that petrologists should, as far as practicable, use an agreed system of nomenclature so that this can become a help rather than a stumbling-block in petrological research. In thinking critically about these matters we have been increasingly impressed by the incongruities and inconsistencies that bedevil petrographic nomenclature. In the nature of things we can do little about it; but for one group of rocks that are much to the fore at the present time (the Central African lavas), something can be done, and the senior author has devised a number of rock-type names based on a new principle which, if approved by petrologists, would simplify existing nomenclature. In regard to classification the requirements of petrologists vary according to their particular interests, which may be mainly chemical or concerned with genetic associations, and doubtless new schemes will be evolved with the growth of knowledge. However, the rocks themselves do not change, and there will always be a need for a scheme of classification and nomenclature such as that which is proposed in this book, based primarily on mineral associations, that is, on the kinds and proportions of the component minerals, and on their textural relationships.

In former editions an early chapter was devoted to an introduction to crystal optics; but textbooks covering this subject are now readily available, and the space formerly occupied thus is now taken up by subject-matter more appropriate to the title of the book. Similarly in the mineralogy section we have concentrated on the important rock-forming minerals with special reference to petrological aspects—their role in rock formation, their interrelationships and, as far as known, the details of their crystallization history.

Since the retirement of the senior author from academic teaching the onus of revision has become increasingly his son's responsibility—a fact duly and gratefully acknowledged by the former.

A. KINGSLEY WELLS.

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INTRODUCTION

THE NATURE OF THE IGNEOUS ROCKS

PETROLOGY is that branch of geological science which has for its special field of study the rocks of the earth's crust. A **rock** is a mineral aggregate: it may consist of one mineral or many; it may be amorphous or crystalline. Its constituent mineral grains may be of such size that they are readily identified on inspection; or they may be so finely crystalline as to be resolvable only in thin section under the microscope. In studying rocks it is necessary to appreciate their field relations—their modes of occurrence and their associates; to identify the minerals present, and to determine their relative abundance; and lastly to comprehend the mutual relationships between the individual components.

It is customary to recognize three main categories of rocks: the **sedimentary**, the **metamorphic**, and the **igneous or eruptive**. The first need no further comment in this volume. The second comprise all those rocks which have been changed in mineral content, in texture, or both, as a consequence of drastic change of environment. The chief agents active in metamorphism are high temperature and great pressure. If the rise of temperature and increase of pressure have been sufficiently drastic, a coarsely crystalline rock may result which will differ from an igneous rock in no obvious respect. The borderline between the metamorphic and the igneous is ill-defined and gradational. We base the term "igneous rocks" on the mass characters, mineral associations and textural quality of rocks which, by direct observation, are known to be products of volcanic activity and which were formed by the consolidation of magma. **Magma**, in turn, is more or less fluid rock-material drawn from different levels in the Earth, energized to the point of being rendered mobile, and in favourable circumstances erupted at the surface as lava. Lava before consolidation *is* magma; but magma is not necessarily wholly liquid. It is a matter again of direct observation that magma often rises to the surface bringing with it (as an essential part of itself) well-formed crystals suspended in liquid. Thus the lava erupted by Vesuvius is sometimes charged with crystals of leucite, while the commonest type of lava, basalt, frequently contains crystals of olivine and augite. In certain circumstances it is inferred that some bodies of magma consist very largely of concentrates of such crystals, lubricated by a minimum amount of liquid, and therefore possessing only limited mobility.

The rocks termed igneous include lavas of a wide range of compositions; but these are closely associated with other rock bodies which consolidated below ground-level and which therefore cannot

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be observed in the process of formation. It is inferred, however, from the facts of mineral composition and textural quality, that these rocks also originated from the same material as the associated lavas. The argument may be developed by working from the known to the unknown: thus basalt forming lava flows is demonstrably magmatic. A widespread type of basalt consists essentially of the minerals olivine, augite and plagioclase. The last two minerals are often involved in a distinctive intergrowth known as ophitic texture. This texture and mineral association are found also in rocks occurring as minor intrusive sheets (dykes and sills). Thin sheets of either kind may consist of basalt identical with that forming the lava flows; but in the thicker ones cooling was slower and a coarser grain was consequently developed. These medium-grained rocks of basaltic composition are termed *dolerites*; but although they are given a distinctive name on account of their different grain-size and mode of occurrence, it is obvious that they were formed from the same material as the lava basalts. Finally, large rock masses (major intrusions) composed of the same minerals, often showing the tell-tale ophitic texture, may be exposed by deep erosion of the roots of a volcanic region. These coarse-grained rocks of basaltic composition are distinguished as *gabbros*. They grade into *dolerites* just as the latter grade into basalts, and there can be no doubt about their close genetic relationship. The three rock-types basalt, *dolerite* and *gabbro* are essentially the same in chemical composition; but, although they are not identical, such differences in texture and in the details of mineral composition as may be noted are readily understandable as arising during cooling in different environments. These three rock-types doubtless had a common origin and are hence termed *comagmatic* (R. A. Daly) or *consanguineous* (A. Harker). The latter term does not stress the magmatic nature of the rocks, but it does emphasize a common origin—they are of the same blood. Many other rocks differing widely from basalts in composition are linked in precisely the same manner as basalt, *dolerite* and *gabbro*. The close genetic relationship is expressed by saying that they belong to the same "clan"—in this particular case the *gabbro* clan. The clan concept is a valuable one and is extensively used in this book.

A body of magma when intruded into the Earth's crust brings with it much heat as well as active volatile substances which react with wall- and roof-rocks causing changes in their composition and structures. Such reconstruction is termed **thermal metamorphism**, and rocks so affected fall in the metamorphic, as distinct from the igneous, category. Usually distinction between the magmatic rocks causing the metamorphism, and the metamorphic rocks resulting from it are clear and unequivocal: the metamorphic effects are

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restricted to a zone surrounding the intrusion and this is often shown on geological maps. The changes are well known and understood. In the deeper regions of the crust the conditions are far different, however: confining pressure and temperatures are both very much higher—the latter must be near-magmatic and much of the material is near-magmatic too. Here the boundaries between igneous and metamorphic become blurred, and it becomes increasingly difficult to distinguish between cause and effect. If subsequently the rocks from such a region are exposed at the surface it may very well be extremely difficult to interpret what is seen. Rocks which have crystallized from a melt will be closely associated with others which were not far removed from the magmatic condition.

Naturally petrologists are interested in the origins of the rocks they study—that is, in **petrogenesis**; but excluding the lavas, we can never *know* with the same degree of certainty how a rock was made, as we can know and understand the facts of its mineral composition and texture. Therefore in this book we stress the observable facts and study the rocks as they are without becoming too deeply involved, we hope, in the more hypothetical problems of ultimate origins.

The omnibus term “igneous rocks” is admittedly not ideal; but it is widely used, and in spite of its shortcomings it is well understood. A new name could doubtless be devised which did not stress connection with fire. S. J. Shand preferred to use the term “eruptive rocks” for those we call “igneous”; but surely an eruption is a breaking-out at the surface: only the extrusive igneous rocks (lavas) are really eruptive. In addition to the latter we are concerned also with the equivalent rocks which were formed in—not on—the crust of the Earth, and which are collectively termed “intrusive,” though this word also is liable to convey a wrong impression. Careful study of field relations in many cases indicates that gabbros, granites, etc., reached the positions they now occupy in the crust by *displacing* pre-existing “country-rock.” These are genuinely intrusive; but in many other cases such evidence is lacking, and field relations suggest, if they do not prove, that the rocks in question originated in place, therefore by *replacement* rather than displacement. This applies more particularly to the rocks rather loosely termed “granites,” concerning which H. H. Read once said: “There are granites and granites,” implying among other things that some granites originate in one way, but others in different ways. The important thing to realize is that in spite of different modes of origin, the rocks themselves are granites, lithologically. We believe strongly that a given rock-name should apply to a given mineral assemblage, irrespective of its origin. In this respect origin is of secondary importance compared with the petrographic (lithological) character of a rock.

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It is upon the latter that systematic classification and rock-nomenclature are based, and anything out of the ordinary (so far as origin is concerned) can be indicated by adding a qualifier to the rock-name. Thus a granite may originate by direct crystallization from a melt; but during a phase of granitization a closely similar (though not necessarily identical) granite may be formed by recrystallization of lava or volcanic "ash" of the right composition in the root region underlying a volcano; while it is well within the realms of possibility that gaseous and fluid substances of magmatic origin rising in advance of the magma itself may so alter certain pre-existing rocks as to form new mineral associations stable under new conditions and again closely resembling, and in some instances indistinguishable from, magmatic granite.

Readers of petrological literature will quickly discover that conflicting opinions on problems of petrogenesis are strongly held, and often forcefully expressed in a manner that appears to be authoritative. Thus one learns from one source that "all granites are magmatic"; but from another that "no granites are magmatic." Actually these are matters of personal opinion and they really do not matter greatly. There is no room for difference of opinion on what granite is, and what it is made of. These are matters of fact which the student can ascertain and check for himself. Our description and discussion of, say, granites must include *all* granites, and this is equally true of all the rocks termed igneous.

PART I

**THE IGNEOUS ROCK-FORMING
MINERALS**

THE MAFIC MINERALS

(1) Introduction: Classification

THE rock-forming minerals may be variously classified, according to the part played by them in the structure and composition of rocks. A useful division is into primary and secondary, the former being further divisible into essential and accessory. (An **essential** mineral is one whose presence is implied in the definition of the rock. An **accessory** mineral is one whose presence or absence does not sensibly affect the character of the rock. Thus, quartz, feldspar and mica are essential constituents of most granites, while zircon, sphene and apatite are accessories.) It should be realized that an accessory mineral in one rock may be an important essential in another. For example, although quartz is an accessory in some gabbros, it is the characteristic mineral of granites, and no rock free from quartz can be termed granite. The accessory minerals are sparsely distributed; but although only one or two may be seen in a hand-specimen, others come to light when thin sections are examined. To make a complete list, samples of the rock must be crushed and the heavier minerals separated by liquids of high specific gravity. This method has been successfully applied in the study of a number of igneous rocks, and has resulted in a considerable advance in our knowledge of the distribution of the more uncommon constituents.¹ Many of these accessory minerals are among the most stable and most durable components of igneous rocks. They thus persist after the essential minerals have been destroyed by weathering, and by their presence add distinctiveness to the sands and other sediments formed by the degradation of igneous rocks. Their detailed study is now an important branch of petrology (see Vol. II of this work).

(Minerals are **secondary** when they have resulted from the alteration or reconstruction of the primary minerals. In altered rocks both essential and accessory minerals may be replaced by secondary ones.)

It should be noted that a mineral may be primary in one rock,

¹ See, for example, Rastall and Wilcockson, *Quart. Journ. Geol. Soc.*, 71 (1915), p. 592.

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but secondary in another; thus primary quartz occurs in granite, while secondary quartz is liberated as a result of alteration of several species of rock-forming minerals.

In the following account, the primary essential minerals of igneous rocks are described first in as much detail as the writers consider necessary for the proper understanding of the igneous rocks in relation to their mineral composition. Particular attention is therefore paid to the internal structure of these minerals, as it is this which controls variation arising from atomic substitution within each "family" of minerals.

The more important secondary minerals and the accessories also receive attention, but their treatment is necessarily selective.

As it is not our intention to attempt to compete with standard textbooks on mineralogy, much detail concerning the physical properties of the minerals has been omitted: such detail, if required, can be readily obtained from one of the standard works of reference.

(2) *Considerations concerning Atomic Structure*

In these chapters the description of most groups of the rock-forming minerals is based on their atomic structure, as revealed by X-ray analysis. The actual technique involved in the latter, and the interpretation of the experimental data are matters for specialists; but the results of their researches are most illuminating to the student of mineralogy, particularly as regards the chemical relationships between allied mineral species and the interdependence between physical characters and the intimate internal atomic structure. We have of necessity drawn largely upon W. L. Bragg's invaluable account of the atomic structure of minerals. One great difficulty arises in connection with using the results of X-ray analysis: it is far from easy to make a useful illustration representing an intricate atomic design in three dimensions. Those which we have introduced must be regarded only as diagrams—they are only first approximations to the truth; but as a teaching device they have their place, if for no other reason than that they help to make sense out of the otherwise meaningless string of symbols which represents the chemical composition of any one of these minerals.

The component atoms or groups of atoms are held together by electrical forces of attraction between oppositely charged bodies of minute size. Those carrying a positive charge are termed **cations**; those with a negative charge are **anions**. For the sake of simplicity we may speak of the size of an atom, meaning the atomic radius, which is the distance, measured in Ångström units, at which mutual repulsion sets in when two atoms approach one another.

In the essential minerals of igneous rocks we are dealing largely with silicates, *i.e.* combinations of silicon (Si) with oxygen (O),

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these being associated with cations in variety. Those commonly occurring include the following, in order of increasing size, the atomic radius being stated in brackets in each case:

Si (0.39)	Na (0.98)
Al (0.57)	Ca (1.06)
Mg (0.78)	K (1.33)
Fe'' (0.83)	

The anions are much less varied. Among them oxygen (O, 1.32) plays the dominant role, but in certain minerals hydroxyl (OH, 1.32) or fluorine (F, 1.33) may replace it to a limited extent. Now all these anions are large compared with the cations; and we may regard the mineral structures as consisting of closely packed anions (chiefly oxygen), with the small cations tucked into the interstices between them. With some of the more complex minerals the number of kinds of atoms may be large, and the atomic design may be

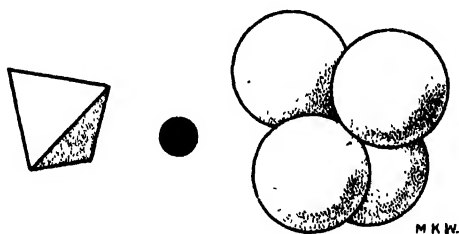


FIG. 1

Tetrahedron and SiO_4 -group.

The small silicon atom is hidden between the four large oxygen atoms, each of which has its centre at a corner of the tetrahedron. The black spot shows a Si-atom to same scale as O-atoms.

exceedingly intricate; but just as an imposing modern skyscraper is built up of relatively insignificant blocks of building material, so these involved atomic structures may be resolved into simple units of structure. For our purposes only two such units need be recognized. In the first, four oxygen atoms are closely packed round a silicon atom, giving an SiO_4 -group. The oxygens lie at the corners of a tetrahedron, and the silicon is of such a size that it fits snugly into the interspace between them, as shown in Fig. 1. The second fundamental unit of structure comprises six oxygens in close contact, lying, as it were, at the corners of an octahedron. There is space between them for a cation larger than a silicon, and atoms of magnesium, aluminium or iron commonly occur in such "six-fold co-ordination" with the anions surrounding them. Each and every cation shares its charge, whether of one, two, three or four units,

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with all the anions by which it is immediately surrounded. Thus an Al''' in the centre of an octahedral group of O's, has a positive charge of three units to share among the surrounding six O's, and therefore contributes a charge of $+\frac{1}{2}$ to each. On the other hand, a silicon (tetravalent with a total charge of $+4$), in an SiO_4 tetrahedron, contributes $+1$ to each of the surrounding oxygens. This is "**Pauling's principle**." In both cases cited, as oxygen is divalent, that is, has a charge of 2 units, it follows that there must be a residual electrical charge on these units of structure. But the whole edifice is compounded of such units, packed together, with interspaces available for the introduction of as many cations as are necessary to balance out this residual negative charge. There must be no residual charge of the kind we have visualized, in a stable mineral. This necessity of balancing the total positive against the total negative charge provides a check on the accuracy of a formula representing the composition of a mineral.

In many rock specimens the component mineral grains are either dark or light in colour, and this simple fact is the basis for dividing the essential rock-forming minerals into two categories "light" and "dark," or more scientifically, mafic and felsic respectively. The mafic group comprises the ferro-magnesian silicates: the olivines, pyroxenes, amphiboles and micas, and these are considered in that order, which is that of increasing structural complexity, in the account which follows. In the following chapter the felsic minerals are described. These include two "families" of anhydrous aluminosilicates, the feldspars and feldspathoids respectively, and the silica group.

OLIVINES

This group comprises a number of important rock-forming silicates. In all members of the group the essential plan of the atomic structure is the same: isolated SiO_4 -tetrahedra are packed together in lines parallel to the crystal axes. In any such line parallel to an axis, they all point in the same direction (Fig. 2); but in alternate lines the tetrahedra point to the left and right, as shown in the spaced diagram of Fig. 3. The individual tetrahedra are joined one to another through the cations. These are arranged in six-fold co-ordination with the oxygen atoms, which, as the diagram shows, belong to different adjacent tetrahedra.

Thus the unit of structure is the single SiO_4 -tetrahedron. Against the negative charge of 8 units supplied by the oxygens, the Si'''' offers a positive charge of 4 units; therefore on each tetrahedron a negative residual charge of 4 units remains to be balanced out by the addition of the requisite cations. In the several members of the

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olivine group the latter include Mg'' , Fe'' , and much less commonly Mn'' and Ca'' . If all the necessary cations are Mg'' , the formula

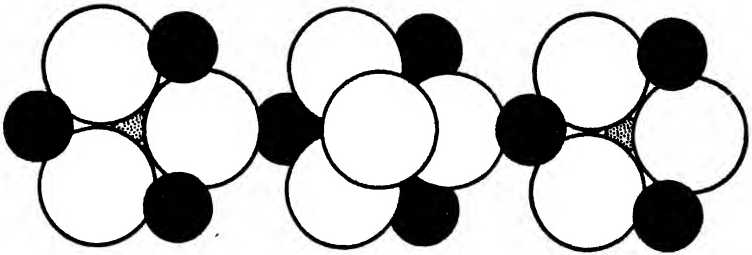


FIG. 2

The atomic structure of olivine

A small portion of a single row of SiO_4 -tetrahedra with the associated cations, Mg and Fe , shown in black, silicons stippled. This row is parallel to the b-axis.

becomes Mg_2SiO_4 , which corresponds to the natural mineral **forsterite**. If iron is used exclusively, the corresponding mineral is **fayalite**, represented by Fe_2SiO_4 . These two orthosilicates, as they

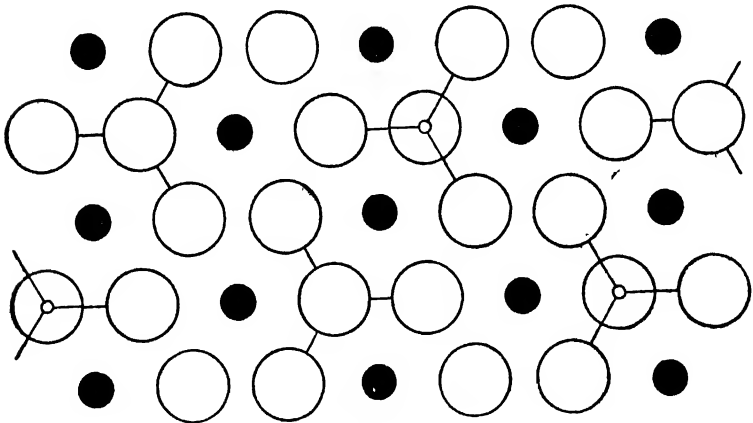


FIG. 3

The atomic structure of olivine represented formally.

Large circles—oxygen; small circles—silicons; black—magnesiums and/or irons. Two rows of SiO_4 -tetrahedra are shown with their oxygens in planes parallel to (100). Each tetrahedron is associated with three metallic cations (Mg'' , Fe''): these three are alternately behind (top row, left), and in front of, the tetrahedra as shown in Fig. 2.

are called, are the end-members of a continuously variable series, in which the ratio of Mg to Fe varies from 100 : 0 to 0 : 100. Such a series may be subdivided arbitrarily into a number of mineral

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species, with agreed ranges of composition. Unfortunately, in this case, agreement has not yet been reached as to how many divisions should be erected, but we show two recent suggestions in Fig. 4. To avoid misconception it is necessary to use a symbol, in addition to, or instead of, the mineral name. The symbol indicates accurately the molecular percentage composition: thus if pure forsterite is represented by Fo and pure fayalite by Fa, the symbol $\text{Fo}_{52}\text{Fa}_{48}$ gives full information as to the composition of this particular specimen. As a matter of interest it may be noted that if names were used, this would be hyalosiderite (Wager and Deer), chrysolite (Tomkeieff), but in the original description of the mineral it was called hortonolite.¹

Although the composition may be expressed in terms of the molecular proportions of the end-members, forsterite and fayalite, this must not be taken to mean that separate molecules occur in the mineral. The silicon-oxygen framework of the atomic structure runs continuously through the crystal, with the Mg'' and Fe'' ions evenly and statistically distributed throughout. It is best to think in terms of ionic substitution in the structure rather than solid solution of one molecular component in another: for this reason there is a growing tendency to quote compositions in terms of the proportions of metallic cations, e.g., $\text{Mg}_{52}\text{Fe}_{48}$.

The stability conditions controlling the crystallization of the members of this series have been studied experimentally.² Under laboratory conditions the temperatures of crystallization range from approximately $1,890^{\circ}\text{C}$. for pure forsterite to $1,205^{\circ}\text{C}$. for pure fayalite (Fig. 4). The diagram shows that for an olivine of composition Fo_{50} , crystallization commences at about $1,650^{\circ}\text{C}$., and that the first-formed crystals are considerably richer in Mg than the original melt.³ As the temperature falls, the composition of the growing crystals changes, becoming progressively richer in Fe. With relatively rapid cooling the successive layers of crystalline material may survive as distinct zones of different composition; the core in such a zoned crystal must be relatively rich in Fo, while the outermost zones must be correspondingly rich in Fa. A difference of the order of 20 per cent Fo has been noted.⁴

Reasoning on similar lines, early-formed phenocrysts in a lava

¹ Wagner, P. A., *Platinum Deposits and Mines of S. Africa*, London, 1929, p. 55.

² Bowen, N. L., and Schairer, J. F., "The System MgO-FeO-SiO_2 ," *Amer. Journ. Sci.*, **29** (1935), p. 151.

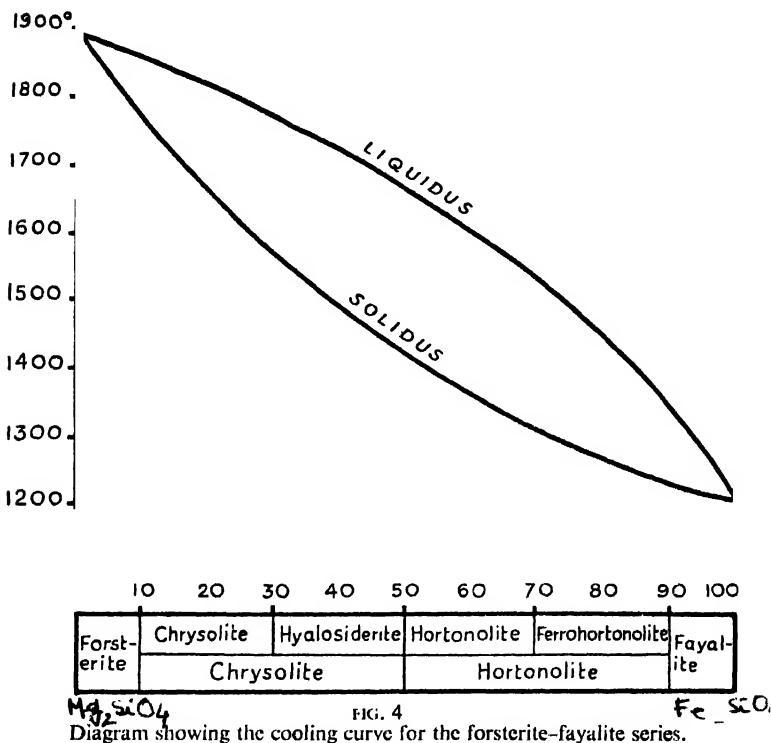
³ Draw a vertical line from 50 to cut the *liquidus* curve, then a horizontal line through this point shows the temperature at which crystallization commences; while the point where the horizontal line cuts the *solidus* curve indicates the composition of the first-formed crystals.

⁴ Tomkeieff, S. I., "Zoned Olivines and their Petrogenetic Significance," *Min. Mag.*, **25** (1939), p. 229.

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should be (and are) richer in Fo than the smaller crystals in the groundmass of the lava. Further, among the latter those of largest size, which presumably started growth early and at a relatively high temperature, are more richly magnesian than smaller ones, of later, lower-temperature formation.

The discovery of these facts concerning the variation of composition in successive crops of olivine crystals, and even within large



After Bowen, N. L., and Schairer, J. F.

The subdivisions of the olivine group as suggested by Wager, L. R., and Deer, W. A. (*Amer. Min.*, 24 (1939), p. 25), and Tomkeieff, S. I. (*Min. Mag.*, 25 (1939), p. 229), are shown above and below respectively.

single olivine crystals, is only possible if means exist for accurately determining the compositions of the specimens under consideration. Obviously careful chemical analysis will give the desired information; but there are certain less costly ways available to the petrologist. In any continuously variable series like the olivines, the physical, including the optical, properties vary systematically with the composition. Much useful data has accumulated from the study

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of isolated olivines, and has been used to construct curves showing these variations (Fig. 5). Two sets of measurements have proved valuable in this connection: (1) the size of the optic axial angle, $2V$ (measured in the mineral), or $2H$ (measured with the Universal stage); and (2) the refractive indices, more especially the mean index β measured with a refractometer. Unfortunately the actual measurement of these properties is a matter for the specialist: they involve methods and apparatus not normally available to students. It is not practicable to determine the composition of an olivine by

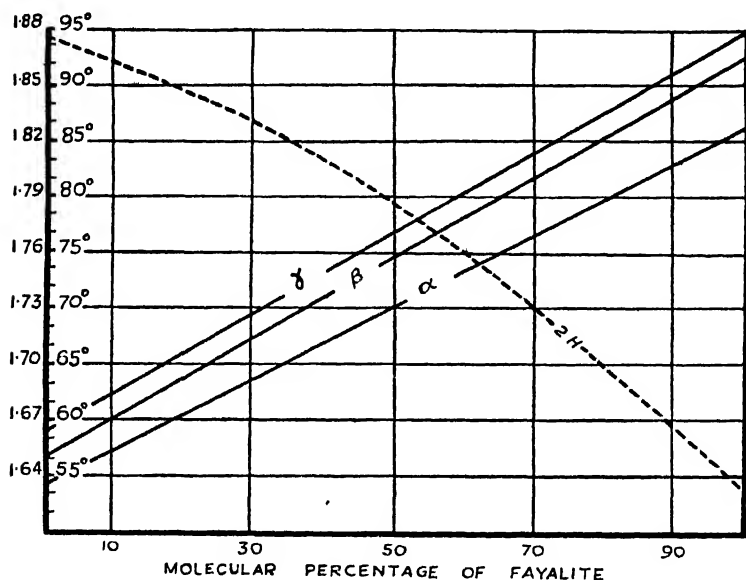


FIG. 5

Diagram showing variation in the size of the apparent optic axial angle $2H$, and refractive indices, in the olivine group. Data from Bowen and Schairer, Wager and Deer, and Tomkeieff.¹

any of the simple methods, such as extinction angle, used in other groups of minerals. While therefore one must be satisfied merely to identify the mineral generically, as it were, it is possible to go a long way towards specific identification in the light of the following facts concerning the distribution of the different members of the group.

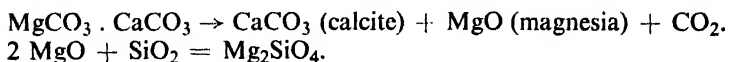
Bowen and Schairer have shown experimentally that Mg-rich olivines—forsterite, chrysolite and hortonolite are unstable in the presence of free silica. Therefore under normal conditions these

¹ In Fig. 5, " $2H$ " is the optic axial angle measured on the universal stage between hemispheres of refractive index 1.649.

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members of the group are not found in association with quartz: no granitic rock can contain Mg-rich olivines. But olivine rich in Fe can exist in the presence of free silica, though even here there is a limitation: the temperature must be relatively low to bring it within the temperature-range of rhyolitic magma. This is merely another way of stating that the only kind of olivine which can occur in a "granitic" association of minerals is nearly pure fayalite. The latter has now been recorded in a number of quartz-bearing rock-types including pitchstones from Arran, Scotland, obsidian from the Yellowstone Park, quartz-porphyrries and rhyolites from Nigeria, etc., but it still remains a relative rarity. The olivines, excluding forsterite and fayalite, are typically components of the silica-poor (basic and ultrabasic) igneous rocks. The amount is greatest in certain ultrabasic rocks, significantly termed olivinites and peridotites,¹ and is only little less important in certain basic igneous rocks such as olivine-gabbros, troctolites, and olivine-basalts. Some of the last-named contain so-called "olivine-nodules," but these are really small pieces of peridotites that were brought up by, and incorporated in, the magma during its uprise towards the surface. Well-known examples occur in the Carboniferous basalts in Derbyshire. Now in all these rocks there is a preponderance of Mg'' over Fe'', and with few exceptions the general rule may be enunciated that the more basic the rock, the more richly magnesian the olivine; the more siliceous the rock, the more ferriferous the olivine.

The end-member—**forsterite**—stands in a category by itself: it is a characteristic product of the thermal metamorphism of magnesian limestones and dolomites which contained the necessary silica in the form of detrital quartz grains, sponge spicules or tests of radiolaria. The double carbonate breaks down into calcite and magnesite, but the latter goes further and yields up CO₂, leaving MgO (magnesia) free to combine with the silica:—



The crystals and grains of forsterite are embedded in a matrix of crystalline calcite, and the rock is termed forsterite-marble.

General Characters of the Group

All members of the olivine group crystallize in the Orthorhombic system; but although olivine is a common mineral, well-formed crystals are rare in ordinary mineral collections. Olivine sands, formed at some points on the coasts of the oceanic volcanic islands, such as Hawaii, consist largely of singularly perfect, though small, crystals, concentrated naturally out of the olivine-basalts which

¹ From "peridot," an alternative name for olivine of gemstone quality.

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form the coastline. Many of these small phenocrysts are tabular, with the pinacoid (100) well developed, combined with the vertical prism (110) and dome (101). Other characteristic crystal habits are illustrated in Fig. 6. Forsteritic olivines are "growth-sensitive": they readily develop different shapes in different environments.

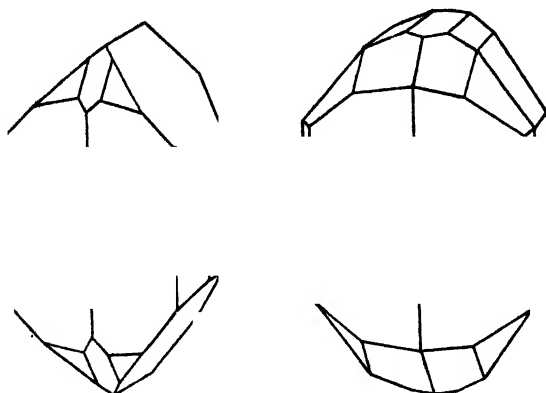


FIG. 6

Crystals of olivine.

Combination of two vertical prisms, side pinacoid, brachyprism {Ok}, macroprism {hOl} (sometimes called domes), and bipyramid. Crystal on right from St. John, Red Sea, shows basal pinacoid in addition, also two brachyprisms and two bipyramids.

On account of the dominance of the tabular habit noted above, in thin sections olivine tends to show six-sided cross sections of characteristic appearance (Fig. 114). In all members of the group the refractive indices are high, and strong surface relief, combined with complete absence of colour, the dominance of arcuate fractures over ill-developed cleavages,¹ and the strong birefringence, serve to render olivine easy to identify. The birefringence of an olivine of known composition may be read off the curve in Fig. 5; that of fayalite is very high. All olivines are prone to alteration in a distinctive manner, and in a representative collection of sections from olivine-basalts, all stages from incipient alteration to complete pseudomorphism may be studied. Commonly fibrous antigorite and/or green chrysotile appear, first along the fractures, then they spread through the body of the crystal. Less ubiquitous secondary minerals formed from olivine include bowlingite and iddingsite, much more strongly coloured "serpentine," yellow to reddish brown in thin section. In other rocks the olivines have been converted into ferruginous pseudomorphs, particularly in basalts

¹ Cleavages develop parallel to the pinacoids (010) and (001) in iron-rich olivines, notably in fayalite.

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weathered under aeolian conditions. In lamprophyres, and occasionally in other rocks, the pseudomorphs may consist largely of carbonate, with or without some form of silica such as opal, chalcedony, or quartz-mosaic. Probably the most carefully studied pseudomorphs after olivine are those occurring in the Markle-type basalts in Scotland. They consist of penninitic chlorite, about 66 per cent, haematite, 32 per cent and 2 per cent of quartz.¹

PYROXENES

The pyroxenes constitute one of the most important groups of rock-forming silicates. Compared with the olivines, they contain a higher proportion of silica to the bases present, and are thus *meta-silicates*. In different members of the group the elements iron, magnesium, calcium and sodium are present in widely varying proportions, together with smaller amounts of aluminium, manganese, and titanium.

Atomic Structure and Chemical Relationships

The study of the X-ray structure of the pyroxenes has shown that the fundamental SiO_4 -tetrahedra are linked together vertically into

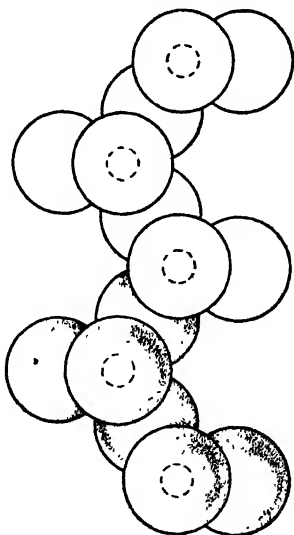


FIG. 7

Part of a chain of SiO_4 -tetrahedra, as in pyroxenes. The unit of pattern contains (Si_2O_6) shown shaded. Si-atoms shown by broken circles; large circles are O-atoms.

¹ Smith, W. W., *Min. Mag.* 32 (1959), p. 324.

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chains, each tetrahedron sharing oxygens with those immediately above and below in the chain. The individual chains are joined together through the medium of the cations Ca'' , Mg'' , Fe'' , etc., which are linked to the "free" (*i.e.* not shared) oxygens. In all pyroxenes the chains run parallel to the vertical crystallographic axis, and are arranged in sheets parallel to (100). The essential features are shown diagrammatically in Figs. 7 and 8.

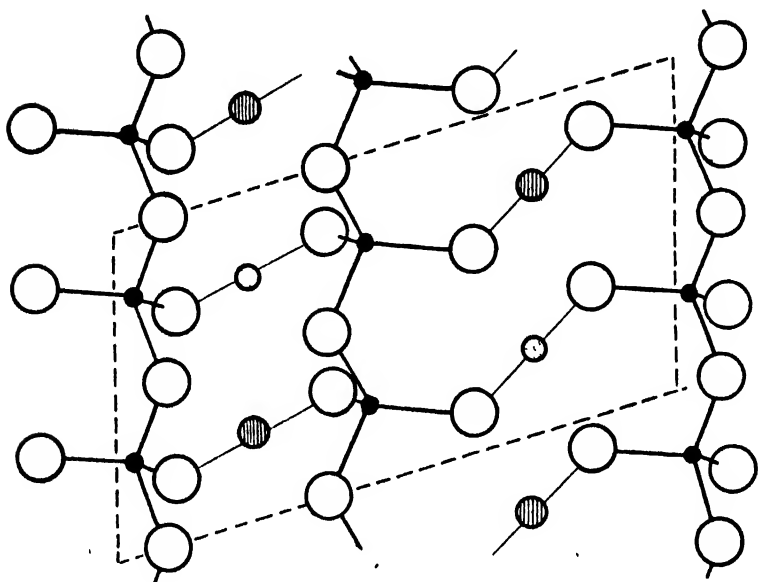


FIG. 8

Diagram showing the structure of diopside.

Portions of three parallel pyroxene chains projected on the plane (010). The unit cell is outlined, but only part of its contents is shown. Si'' black; O'' plain; Mg'' stippled; Ca'' ruled.

The form and disposition of the chains determines the positions of the cleavage planes (Fig. 9). These are parallel to the prism faces of the pyroxene crystal, and cross at a characteristic angle of 87° (or 93°).

The unit of pattern in the pyroxene chain contains Si_2O_6 . To balance the valency two Mg'' ions may be added, giving $\text{Mg}_2\text{Si}_2\text{O}_6$, which, of course, may be cancelled down to MgSiO_3 —magnesium metasilicate—or the mineral *enstatite*. In *ferrosilite* all the cations are Fe'' , the formula being FeSiO_3 . Between these two extremes every gradation exists. A third essential component of the pyroxenes is the corresponding metasilicate of calcium, CaSiO_3 , which

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is the formula of the mineral wollastonite.¹ Now ignoring certain minor constituents, the composition of the pyroxenes may be represented by points on a triangle (Fig. 10). The apices of the triangle represent the pure metasilicates of calcium, represented by

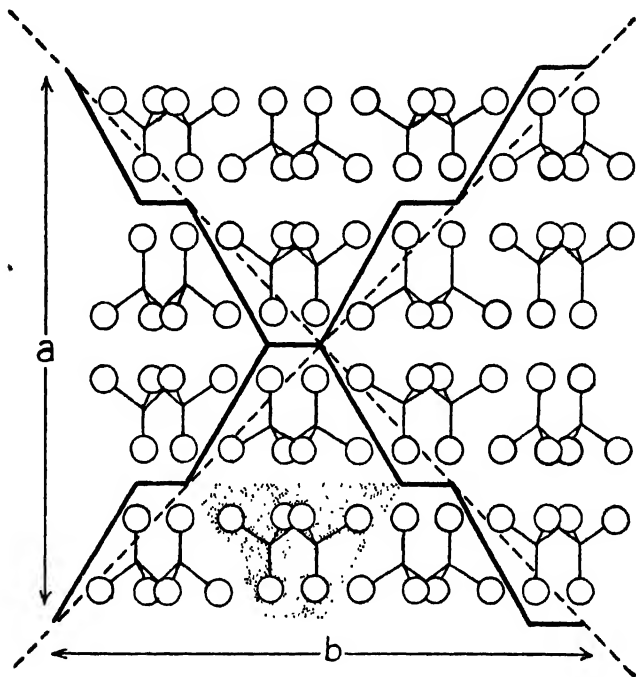


FIG. 9

Diagram showing the relationship between the atomic structure and cleavage of pyroxenes. The linked Si-O chains are shown in plan and the position occupied by one chain with its cations (not shown) is stippled. The planes of weakness are shown by heavy lines, and the resulting cleavage directions by broken lines: *a* and *b* axes shown; *c* is perpendicular to the paper.

Wo, of magnesium, written En, and of iron, written Fs. All pyroxenes may be represented by a simple formula expressing the molecular percentages of these three components. The base of the triangle embraces all possible proportions of En to Fs, thus covering the compositional range of the important group of orthopyroxenes, and the chemically identical monoclinic equivalents. The naturally

¹ Although calcium metasilicate, CaSiO_3 , here represented by Wo, is an essential component of the more complex pyroxenes, the mineral wollastonite, with the same formula, is not a pyroxene, as it has a different atomic structure. It is not a normal constituent of igneous rocks, but is characteristic of the thermal alteration of limestones. As wollastonite is like the pyroxenes, though not isomorphous with them, it has been termed a "pyroxenoid."

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occurring members of the enstatite-ferrosilite series contain a small amount, usually between 4 and 5 per cent, of the Wo component.

The mineral **diopside** (Di), with the formula $\text{CaMgSi}_2\text{O}_6$, is represented by the point half-way between Wo and En; and similarly the analogous silicate **hedenbergite** (He), $\text{CaFeSi}_2\text{O}_6$, lies half-way between Wo and Fs. A continuously variable series links Di with He.

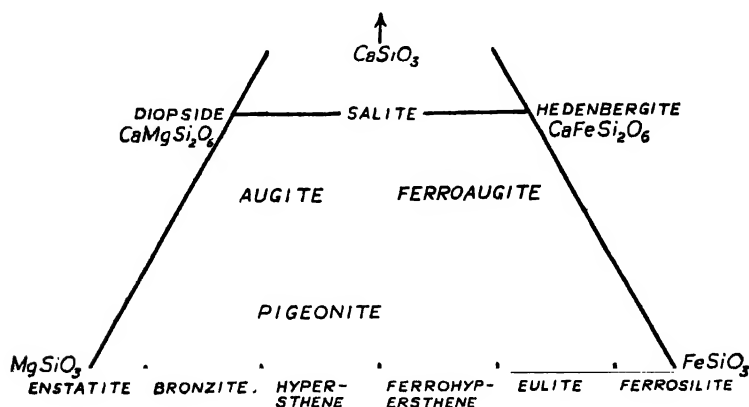


FIG. 10

Triangular diagram showing the compositional relationships between the common pyroxenes.

No pyroxenes lie above the Di-He line; but there remain some of the most important pyroxenes, which in theory go far towards bridging the gap between the diopside-hedenbergite series on the one hand, and the enstatite-ferrosilite series on the other. These are grouped as the pigeonites and the augites—both clinopyroxenes, but distinct mineral species. Pigeonites are calcium-poor; they are related in composition to the orthopyroxenes, but differ from them in containing more of the Wo component. H. H. Hess suggests a range of composition from Wo_5 to Wo_{15} .

The augites lie between the pigeonites and the diopside-hedenbergite series. The most important component not shown on the triangle is alumina, a moderate amount of which is present in both augites and pigeonites.

Thus from the diagram it is seen that there are four series of pyroxenes to be considered:

- (a) the enstatite-ferrosilite series known collectively as the orthopyroxenes, as they crystallize in the Orthorhombic System;
- (b) the pigeonite series, restricted to rapidly cooled basic igneous rocks, especially to some varieties of basalt;

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- (c) the important and widely distributed augite-ferroaugite series; and
- (d) the diopside-hedenbergite series, rare in igneous rocks, but common in metamorphic rocks.

In the following account the pyroxenes are described in the above order.

(a) ORTHOPYROXENES: THE ENSTATITE-FERROSILITE SERIES

Minerals ranging in composition between MgSiO_3 and FeSiO_3 comprise two series, one crystallizing in the Monoclinic System and distinguished as clino-enstatite, clino-hypersthene, etc. These minerals are stable under high-temperature conditions only: they do not occur in igneous rocks but are encountered in meteorites and as products of dry-melt experiments. The second series is by far the more important; these pyroxenes crystallize in the Orthorhombic System and are therefore distinguished as orthopyroxenes. They are widely distributed in a large variety of igneous rock-types. The relationship between these two series has been established experimentally by Bowen and Schairer, who demonstrated that the Monoclinic members invert into their Orthorhombic equivalents at temperatures of the order of $1,000^\circ \text{C}$.¹

For convenience the isomorphous series of orthopyroxenes is subdivided into six "species" defined in terms of the molecular percentages of MgSiO_3 and FeSiO_3 , the divisions between the ranges of composition being drawn at 10, 30, 50, 70 and 90 mols. per cent, as is the common practice for the important groups of the olivines and plagioclase feldspars.

Crystals of orthopyroxenes are not normally available for study; and the student's knowledge of these minerals is based on the examination of cleavage masses or aggregates of hypersthene or bronzite, bounded in part by lustrous bronze-coloured cleavage-surfaces. However, in thin sections particularly of certain lavas, the characteristic shapes of euhedral phenocrysts give the first indication of the presence of orthopyroxene. The most distinctive are basal (transverse) sections, nearly square in shape with the corners truncated, cut across stumpy prismatic crystals. These sections also show well-developed prismatic cleavages, and often in addition an extra single pinacoidal cleavage. The sections show moderate relief: they are either colourless or faintly coloured, and equally faintly pleochroic, so that this feature may easily be overlooked. In thicker sections the pleochroism is very distinctive from deep pink to bluish green. Enstatite is colourless in thin section, but hypersthene is

¹ "The System MgO-FeO-SiO_2 ," *Amer. Journ. Sci.*, **29** (1935), p. 151.

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faintly coloured and feebly pleochroic. Both colour and pleochroism become more intense with increasing Fe-content. The birefringence of orthopyroxenes is weak and the interference colours are limited to shades of grey. Extinction varies, of course, with the orientation of the section. The student should not expect that all sections of these Orthorhombic minerals will show straight extinction: only those sections cut parallel to a crystallographic axis will do so—other sections may extinguish at a wide angle. Similarly the fact that one or two pyroxene sections in a slide show a grey interference colour does not necessarily prove that they are orthopyroxenes: certain sections of clinopyroxene (augite for example) will show this if they are cut perpendicular to an optic axis. Orthopyroxene does not show twinning effects unless it has inverted from pigeonite (see below).

A special feature characteristic of orthopyroxenes occurring in many coarse-grained igneous rocks is the intergrowth with a second (clino-) pyroxene of contrasted composition. The latter is enclosed in the host mineral as regularly orientated plates, in some instances excessively thin, of diopsidic augite, whose identity is easily established by its stronger birefringence as compared with the orthopyroxene host. The theoretical significance of this phenomenon is considered below under exsolution in pyroxenes.

The role of orthopyroxenes as rock-formers is an important one: they occur in lavas, dyke-rocks and plutonites of widely varying composition—from thoroughly Acid to Ultrabasic. This is true of enstatite, hypersthene and bronzite; but iron-rich varieties are less widely distributed and some are very rare. Thus eulite occurs only in rocks of rather problematic origin termed eulysites; ferrosilite is very rare indeed but has been recorded from small steam-cavities in rhyolitic obsidian.¹

Among occurrences in lavas, andesites provide the most satisfactory sections for studying enstatite and hypersthene (Fig. 105). Among medium-grained igneous rocks certain quartz-dolerites, including the well-known British Whin Sill, contain hypersthene associated with other pyroxenes. In the coarse-grained category orthopyroxenes are most important in Basic and Ultramafic rocks, culminating in the virtually monomineralic bronzitite; the two-mineral rock bronzite-peridotite is another significant type; but the commonest Basic rock in this category is norite in which the dominant pyroxene is hypersthene or bronzite.

The stability relationships between the several pyroxenes can only be understood when all the pyroxenes have been reviewed: these matters are considered at the end of this account.

¹ Bowen, N. L., "Ferrosilite as a natural mineral," *Amer. Journ. Sci.*, **29** (1935), p. 151.

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(b) THE PIGEONITE SERIES

Pigeonites take their name from Pigeon Point, Minnesota. They were formerly named enstatite-augite which indicates the essential feature of their chemical composition: they are in this sense intermediate between the augites and the orthopyroxenes. From the position of the word "pigeonite" on the diagram (Fig. 10), it will be noted that "hypersthene-augite" would have conveyed a truer indication of the chemical composition of the pigeonites.

Pigeonites are Monoclinic, but the crystallographic characters are unimportant, as crystals are not available for study. The field of occurrence is much more limited than in the case of the orthopyroxenes, as the pigeonites are restricted to lavas within a relatively narrow range of composition: they are found to occur in basalts only; in fact the presence of pigeonite in certain basalts has led to the recognition of a pigeonite series of basalts (see "Classification of Basalts," Chapter VII. A certain amount of difficulty arises from the fact that pigeonite hardly ever occurs alone in these rocks: it is usually associated with augite, and sometimes with augite and orthopyroxene, so that accurate identification is essential.

Pigeonite very closely resembles augite in thin section in regard to its general optical characters, but differs in the orientation of the optic axial plane and in the size of the optic axial angle. Many pigeonites are uniaxial and the others are only just biaxial, so that identification depends upon establishing this fact. The observer should concentrate on grey-polarizing sections—the deeper the grey, the better—and examine these in convergent light arranged for showing interference figures. Once the identity of one grain has been established by this test, very careful comparison will probably enable one to see other pigeonite crystal-grains and to form an estimate of its relative importance in the rock-section.

(c) THE AUGITE-FERROAUGITE SERIES

In one form or another augite is by far the best known pyroxene. The familiar brown-black crystals similar to that illustrated in Fig. 11 are among the best known Monoclinic crystals available for study. Common augite is characteristic of Basic igneous rocks and occurs in countless basalts, dolerites and gabbros. The average composition is represented by the molecular formula $\text{Wo}_{40}\text{En}_{45}\text{Fs}_{15}$. Augites which crystallize early during the fractionation of a basic magma tend to be Mg-rich; but those which separate later become progressively richer in ferrosilite and are distinguished as **ferro-augite**.

Two other variants of augite should be noted. The first is titanaugite, obviously so called on account of its appreciable content

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of titanium; the second is aegirine-augite, a variety which contains a proportion of the sodic aegirine molecule (see below).

Titanaugite is readily identified under the microscope: it resembles common augite in all respects but one—it is distinctly pleochroic from lilac to light brown. Titanaugite occurs in alkali-rich basic igneous rocks, for example in the Scottish essexites and lugarite in which its optical characters are very well displayed.

Aegirine-augite is the most strongly coloured augite: it may be rich green in a slide of normal thickness, and is distinctly pleochroic. In these respects it is much closer to aegirine than to common augite; but is distinguished therefrom by its wide extinction angle.

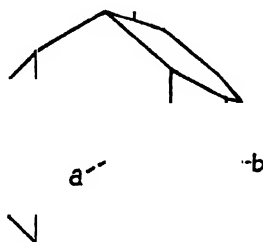


FIG. 11

Augite crystal showing the following forms: front pinacoid (100), vertical prism (110), side pinacoid (010) and hemibipyramid (111).

Note the orientation of the crystal relative to the crystal axes *a*, *b* and *c*. The *a*-axis slopes downwards towards the observer; but the edge between the top hemibipyramid faces (cut by the *c*-axis) slopes downwards away from the observer.

Aegirine and Acmite.¹ These two Monoclinic pyroxenes, occurring in Na-rich igneous rocks, fill the role of augite in calc-alkaline gabbros and basalts. A further link with augite is afforded by aegirine-augite, which, as the hyphenated name suggests, is intermediate in composition and physical (including optical) character between the two pyroxenes named. A common associate of aegirine is nepheline, which is rather elusive in thin sections and may be easily overlooked. The identification of aegirine, which is very distinctive, is a useful hint, therefore, that the presence of nepheline may be suspected. Aegirine is commonly associated, especially in coarse-grained rocks, with an equally strongly coloured sodic amphibole, often in reaction relationship. These latter are very strongly coloured and often violently pleochroic: it is instructive

¹ Sabine, P. A., *Min. Mag.*, 29 (1950), p. 113.

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to compare the optical behaviour of these two kinds of mineral, particularly as regards pleochroism.

Aegirine and acmite are the same in chemical composition, $\text{NaFe}^{++}\text{Si}_2\text{O}_6$; both occur as black lustrous narrow prisms, sharply pointed in the case of acmite. The essential difference in thin section is the tendency for acmite to be brown rather than green.

Jadeite can most conveniently be considered at this point as it has the same type of formula as aegirine (acmite) but with Fe^{++} completely replaced by Al^{+++} . Thus the formula is $\text{NaAlSi}_2\text{O}_6$. As regards the formula, comparison should be made with nepheline on the one hand and albite on the other. In some of the rare occurrences of jadeite—they can be counted on the fingers of one hand—the mineral has resulted from the desilication of albite: at Kotaki in Japan, for example, “lenses” of albitite are embedded in serpentine and are encased in jadeite. The jadeite molecule, in a manner of speaking, enters into the composition of most augites, usually to a very small extent, indicated by the presence of Na' as a minor component; but it is much more important in the pyroxene omphacite which is one of the two essential minerals in the interesting rock-type, eclogite. This rock is of particularly deep-seated origin and it is inferred, therefore, that omphacite can crystallize only in environments involving exceptionally high pressures. This is presumably why jadeite itself is so rare.

In the massive form jadeite, popularly termed “jade,” is well known and highly prized especially in the Far East, as gemstone material obtained from localities in Burma and Tibet.

(d) THE DIOPSIDE-HEDENBERGITE SERIES

Ideally the members of this series are represented by points on the left-to-right join half-way between base and apex of the compositional triangle of Fig. 10; actually the compositions of natural specimens lie a little below this line and are gradational into the augites—an arbitrary division between the two series must therefore be drawn, and we follow H. H. Hess in selecting Wo_{45} as the critical composition. This is the least amount of the wollastonite component that diopside may contain.¹

The Mg-rich members of the series are by far the commonest. Diopside itself is essentially a metamorphic mineral; but diopsidic augite is not uncommon in igneous rocks including some pegmatites. Chrome-diopside and the closely related chrome-augite may be considered together: as the names imply these clinopyroxenes only differ from diopside and augite in containing a small quantity of Cr_2O_3 , which results in a rich green colour both in hand-specimens

¹ Hess, H. H., “Pyroxenes of the common mafic magmas,” *Amer. Min.*, 26 (1941), pp. 515–35; also *Amer. Min.*, 34 (1949), p. 621.

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and thin sections. These minerals are of very limited distribution in igneous rocks, being restricted to certain peridotites.

Hedenbergite is a rare mineral, but has been recorded from porphyritic microgranites in ring-complexes in northern Nigeria and in a granophyre forming part of the Skaergaard Complex in Greenland.¹ Apparently it is only in very late fractions that the high concentration of iron essential to the formation of hedenbergite is achieved.

The Optical Characters of the Clinopyroxenes

The plane of optical symmetry coincides with the single crystallographic plane of symmetry, parallel to (010) in the clinopyroxenes (Fig. 12), except in some pigeonites in which it is at right-angles to

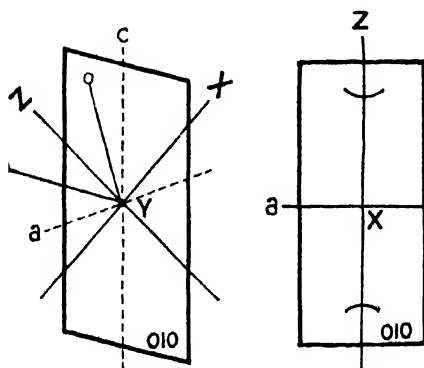


FIG. 12 *Left*, the optical orientation of augite.

The (010) section illustrated is the single vertical plane of symmetry and coincides with the optic axial plane. Note the positions of the optic axes (small circles), and also the extinction angle between Z and c .

FIG. 12 *Right*, the optical orientation of hypersthene.

X (the fast vibration), coincides with the b -axis, Y with the a -axis and Z (the slow vibration) with the c -axis. Positions of the optic axes shown conventionally by small arcs.

that plane. It follows, therefore, that the maximum extinction angle can be measured in (010) sections. These angles vary with composition and provide an easy means of checking the identity of the various clinopyroxenes as indicated in Fig. 13. The angles indicated in the left top corners of the "fans" to the left of the middle line are measured between the cleavage traces and the slow vibration, Z . The complementary angles, between the fast vibration direction, X , and the cleavage traces are obtained by subtraction from 90° . The results would be anomalous if the nature of the vibration under test

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(whether fast or slow) were ignored; and further, it is only the *maximum* angle that is diagnostic. A section parallel to the front pinacoid (100) of augite extinguishes straight (*i.e.* extinction angle,

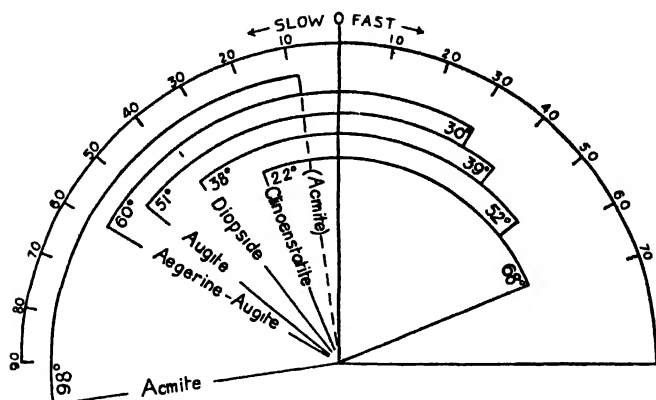


FIG. 13

Diagram of the extinction angles in monoclinic pyroxenes in sections parallel to the side-pinacoid (010). The vertical line (O) represents the vertical cross-wire; the sides of the fans correspond to the cleavage traces on (010) sections in the extinction positions.

0°). The maximum, measured on (010), is 51; therefore sections in the zone (100) (010) show angles between these two extremes. Various sections of augite, drawn in the position of extinction, are shown in Fig. 14. As regards birefringence the range of interference

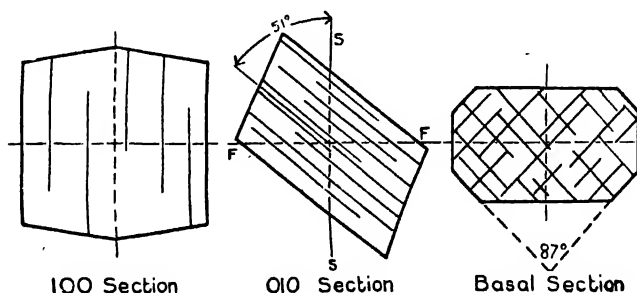


FIG. 14

Sections through an augite crystal drawn in the extinction positions. SS—slow vibration; FF—fast vibration direction. The (100) section extinguishes straight; the basal section, symmetrically; and the (010) section obliquely at the maximum angle.

colours appropriate to the top of the first order and bottom of the second order ensures rich colouring between crossed polarizers, strong yellow, blue, green, red and purple being characteristic. As

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pointed out above, certain sections polarize in dark grey: these should be looked at carefully as regards colour, relief and degree of alteration, when it will be realized that such sections are of the same mineral, *i.e.* augite, as the more brilliantly polarizing ones. As most clinopyroxenes are members of continuously variable series, they are liable to be zoned: certain titanaugites in particular display this feature to advantage. Hour-glass structure is a kindred phenomenon resulting perhaps from selective adsorption of ions on different faces of growing crystals. The two effects in combination are illustrated in Fig. 15.

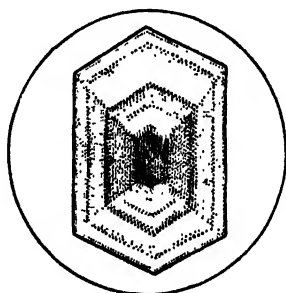


FIG. 15

Zoning and hour-glass structure in augite in lamprophyre, Mühlörzen, Mittelgebirge, Bohemia.

Twinning, either simple or repeated, is commonly exhibited by augite. The twin plane is (100), *i.e.* coincident with the left-to-right cross-wire in the basal section shown in Fig. 14. More complicated twins, commonly cruciform, are often seen, especially in sections of basic lavas.

Crystallization of Pyroxenes

In all four series of pyroxenes falling temperature during crystallization causes a shift of composition from left to right on the composition diagram (Fig. 10), *i.e.* from Mg''-rich to Fe''-rich types. This progressive iron-enrichment manifests itself in three ways: (1) crystals may be zoned, with Mg''-rich cores and Fe''-rich outer layers; (2) early-formed phenocrysts are more magnesian than the crystal grains or microlites occurring in the groundmass of a lava or dyke-rock; (3) in a series of associated rocks representing successive magma-fractions, pyroxenes from early fractions are more magnesian than later ones.

The Monoclinic diopside type of structure is stable over a wide temperature range, and therefore augite is ubiquitous in Basic igneous rocks throughout the whole range of grain-size variation, and is common also in rocks of Intermediate composition, especially

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in andesitic lavas. Pigeonites and orthopyroxenes are less widely distributed for two reasons: firstly, the former are stable only at relatively high temperatures and invert into the latter on cooling; and secondly, MgFe-rich pyroxenes must be regarded as alternatives to olivines in circumstances explained below.

The temperature of inversion from pigeonite to orthopyroxene falls from 1,140° C. for the Mg'' end member of the series to 980° C. for an iron-rich orthopyroxene, say Fs₈₀. The temperature at which pyroxene begins to crystallize from a magma depends upon the composition of the latter. In Mg-rich basic magmas pyroxene commences to crystallize at temperatures slightly below the clinopyroxene inversion, so that the first precipitated pyroxenes are enstatites and bronzitites. On the other hand iron-rich later magma fractions begin to crystallize at temperatures *above* the inversion point, so that pigeonite (usually within the hypersthene range of composition) is first precipitated. In other words the curve representing magmatic temperatures plotted against the composition of the pyroxenes may intersect that of inversion temperatures plotted on the same basis (the Mg : Fe ratio).

It will be apparent from this discussion that careful study of the pyroxenes in a rock may indicate the temperature at which it crystallized. The technique was originally suggested by H. H. Hess,¹ who successfully applied it to the Stillwater Complex in Montana. The significant observation is the first appearance of pigeonite in the rock concerned, since this marks the point in fractional crystallization where the curves intersect. Composition of the pigeonite is determined and the point corresponding with this composition on the inversion curve indicates the magmatic temperature. In the case cited the composition of the first-formed pigeonite was found to be En₇₀Fs₃₀, corresponding to a temperature of 1,100° C. What happens subsequently depends upon the rate of cooling: if the temperature falls fast, as in the case of a lava, pigeonite may survive in the metastable condition; but with slower cooling, under plutonic conditions, it will invert into orthopyroxene. Such "inverted pigeonites" may be easily recognized by the inclusion of relatively thick lamellae of exsolved diopsidic clinopyroxene, lying parallel to the (001) plane of the original pigeonite (see below).

Exsolution phenomena in pyroxenes

Under conditions involving slow cooling, pyroxenes acquire distinctive structural features not exhibited by their high-temperature equivalents. Firstly, pigeonite cannot survive slow cooling but inverts into orthopyroxene. As indicated on the composition diagram, pigeonite contains a good deal less Mg'' and Fe'' than an

¹ "Pyroxenes of common mafic magmas," *Amer. Min.*, 26 (1941), p. 573.

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orthopyroxene does, but is correspondingly richer in Ca'' : the former contains about 9.5 per cent CaO , as compared with 1.6 in the latter. Therefore, when pigeonite inverts into orthopyroxene there is a large surplus of CaO which cannot be accommodated within the structure of the latter. It is therefore **exsolved**. On account of limited ionic diffusion through solid crystals the unwanted material cannot be completely eliminated, but it is concentrated in parallel, minutely thin lamellae which have the composition of diopsidic clinopyroxene. Two different structures of this kind have been described: the first, named the Bushveld type from its typical occurrence in the complex of that name, involves orthopyroxene as the host mineral, containing exsolved lamellae, excessively thin—from 100 to 250 per m μ . of thickness—and parallel to the (100) plane (Fig. 70). The second type is named from its typical development in the Stillwater Complex, and differs from the Bushveld type in two respects: the lamellae are thicker (25 to 100 times thicker) and they are parallel to the monoclinic plane (001). Hess has suggested that the coarse (001) lamination resulted from exsolution taking place mainly during the inversion; while the fine (100) lamination was effected with greater difficulty at a temperature *below* the inversion point. When the (001) lamination described above is combined with twinning on (100), the parallel lamellae slope away from the twin plane in a very distinctive manner, giving a "herring-bone" structure. This proves the orthopyroxene to be inverted pigeonite, for twinning on (100)—the common type of augite twin—although appropriate to Monoclinic minerals, including pigeonite, is impossible in orthopyroxene.

One other type of exsolution structure is sometimes encountered in dolerites—rocks of the same composition as the coarse-grained gabbros and norites occurring in the Bushveld and Stillwater Complexes, but of medium grain, indicating more rapid cooling. Under these conditions the exsolved clinopyroxene takes the form of small *grains* of diopside disseminated throughout the orthopyroxene instead of being concentrated into structurally aligned lamellae.

Augites in some basic intrusives also show exsolution interlamination exactly comparable with those just described, but with the roles reversed: the host mineral is augite, but the exsolved lamellae are orthopyroxene. Presumably this is a direct consequence of the change in composition of augites with falling temperature. Augites of early, high temperature formation are richer in MgFe'' and correspondingly poorer in Ca'' than those which crystallized later, at lower temperatures. The change in composition involves not only iron-enrichment as noted above, but also enrichment in Ca'' , so that the compositional change moves towards the right

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and upwards towards the diopside-hedenbergite join in Fig. 10. Early augites contain about Ca_{40} and late augites Ca_{50} per cent of the total metallic cations.

Relationship between olivines and pyroxenes

It was pointed out above that a reaction relationship exists between olivines and orthopyroxenes. The former crystallize at higher temperatures than the latter; but early formed olivine may react with the magma by extracting silica from it in sufficient quantity to convert it into the corresponding orthopyroxene. The conversion requires time for its completion, however, and in the case of a lava, cooling may be so rapid that the reaction temperature is passed over too quickly for it to be effected, and some olivine may survive, even when the magma contains sufficient silica to convert it all into orthopyroxene. Normally in the more highly silicated magmas, for example those of andesitic composition, olivine does not occur, though orthopyroxenes, including enstatite and hypersthene, may be well represented and give no indication of having "evolved" from earlier olivines (see Fig. 105). The absence of pigeonite from these lavas is a direct consequence of their crystallization from cooler magmas below the pigeonite-orthopyroxene inversion temperature.

Basaltic magmas on the other hand are relatively hot, temperatures of the order of $1,100^{\circ}\text{C}$. having been measured in the basalt "lava lakes" in Hawaii. They are also more basic and therefore olivine is inevitable unless the magma is unusually siliceous for a basalt. The early crystallization of this mineral withdraws much MgFe from the magma and this naturally affects the composition of the pyroxene which is precipitated in due course: it is a Ca'' -rich clinopyroxene (augite) which occurs in association with abundant olivine in many olivine-basalts. Two essential ions in augite are Ca'' and Al''' ; but these are essential too in the plagioclase which accompanies olivine and pyroxenes in these rocks, and which in order to simplify matters we will regard as having the composition of anorthite, $\text{CaAl}_2\text{Si}_2\text{O}_8$. The early separation of anorthite in quantity will deplete the magma in $\text{Ca}''\text{Al}'''$, necessary for the formation of augite, and therefore this may be suppressed in favour of bronzite or hypersthene. Thus olivine and anorthite work in opposite directions and what exactly happens in any specific case depends upon several factors which it would be unprofitable to examine further at this stage.

AMPHIBOLES

The amphiboles form a large group of complex metasilicates, and are chemically related to the pyroxenes. Any species of pyroxene

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may contain identically the same elements as the corresponding amphibole, but they are present in different proportions; while a more fundamental difference is the presence in the latter of hydroxyl, represented by (OH), and with a negative charge of one unit. A hydroxyl group is the same size, and functions in the same way, as an oxygen atom.

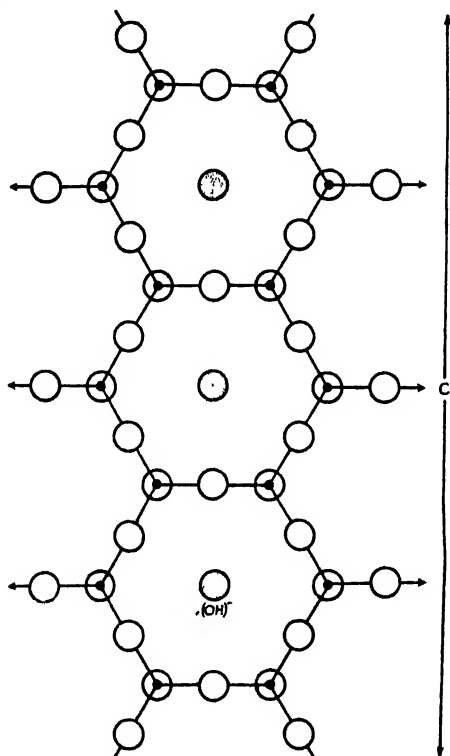


FIG. 16

Diagram of the atomic structure of the amphiboles: the amphibole "band." The arrows indicate the linkage of "free" oxygen atoms to cations. The oxygens superimposed on the Si atoms are similarly linked to cations, thus isolating each Si-O band from its neighbours.

(1) Chemical Relationships

The X-ray structure consists fundamentally of "bands" of linked SiO_4 -tetrahedra. Each band in effect consists of two pyroxene chains united by shared oxygens as shown in Figs. 16 and 17. The arrangement of the bands parallel to the vertical axis is essentially the same as for the pyroxene chains, illustrated in Fig. 7.

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The unit of pattern in Fig. 16 contains Si_4O_{11} ; but to avoid the necessity of splitting an atom, it is convenient to double this. Each of these Si_8O_{22} units is associated with two hydroxyls (OH). Thus the total negative electrostatic charge on the unit is 46, while this is partly counterbalanced by the positive charge which amounts to 32, due to the silicons. The negative residual charge of 14 units must be balanced out by the addition of the requisite number of cations, usually seven. The latter are the same as those occurring in the pyroxenes, *i.e.* Mg'' , Fe'' , Fe''' , Na' and Ca'' . If all the necessary cations are Mg'' , the formula becomes $(\text{OH})_2\text{Mg}_7\text{Si}_8\text{O}_{22}$, which represents the composition of the amphibole **kupferite** equivalent

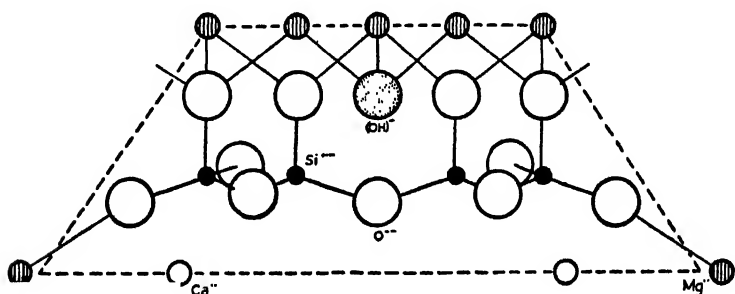


FIG. 17

Diagram of end view of the amphibole band. B-axis, left to right; C-axis at right angles to the paper. The two Mg atoms outside the boundary of the band belong to adjacent bands immediately left and right of the one illustrated. Only some of the bonds are represented.

to enstatite. Similarly $(\text{OH})_2\text{Fe}_7\text{Si}_8\text{O}_{22}$ is grunerite, a monoclinic amphibole formed under metamorphic conditions, and equivalent to the pyroxene clinoferrosilite. These two end-members are connected by **anthophyllite** (orthohombic) and **cummingtonite** (monoclinic), with a variable ratio of Mg to Fe, and thus corresponding to the En-Fs range of pyroxenes (bronzite, hypersthene and eulite). It has not been found necessary to subdivide the Mg-Fe amphiboles in the same way as the pyroxenes, as they have not the same importance in igneous rocks as the latter. The amphibole equivalent to diopside is **tremolite**, which contains both calcium and magnesium, in the proportion of 2 to 5: thus— $(\text{OH})_2\text{Ca}_2\text{Mg}_5\text{Si}_8\text{O}_{22}$. As Fe'' replaces some of the Mg'' in this formula, tremolite gives place to actinolite, which is thus similar to the salite-hedenbergite range.

The commonest, and by far the most widespread, amphibole is hornblende, the equivalent of augite among the pyroxenes. Like the latter, **hornblende** is of extremely complex composition as indicated by the formula: $(\text{Ca}, \text{Na})_2(\text{Na}, \text{K})_{0-1}\text{Mg}(\text{Mg}, \text{Al})_4(\text{Al}, \text{Si}_2)\text{Si}_6\text{O}_{22}$

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(O,OH,F)₂, with Fe'' replacing Mg'', and Fe''' replacing the Al''' outside the chains.

If, in the formula for tremolite, we substitute Na'Fe''' for Ca''Mg'', the amphibole, riebeckite, is derived. This is closely comparable with aegirine and is the most widely distributed of the sodic amphiboles. The formula may be written $\text{Na}_2\text{Fe}''_2\text{Fe}'''_3\text{Si}_8\text{O}_{22}(\text{OH})_2$. Similarly in the latter, Mg''Al''' may be substituted for Fe''Fe''', giving glaucophane, $\text{Na}_2\text{Al}_2\text{Mg}_3\text{Si}_8\text{O}_{22}(\text{OH})_2$. Glaucophane grades into riebeckite, and natural specimens usually have a composition between the two extremes. Allied closely to these amphiboles are barkevikite, arfvedsonite and kataphorite, all three types rich in Na' and ferrous iron.

The ideal formula for **kataphorite** may be written thus: $(\text{OH})_2\text{Na}_2\text{CaFe}_4\text{Fe}(\text{Si}_7\text{Al})\text{O}_{22}$. It will be noted that instead of the Si₈ of other amphibole formulae, that of kataphorite contains Si₇Al. This Al'''-ion is in tetrahedral (four-fold) co-ordination like the silicons in the structure, instead of the usual six-fold, octahedral, co-ordination. It functions in the structure like silicon and is therefore termed a "proxy-aluminium" ion. The substitution of Al''' for Si''' upsets the valency balance, of course, but this is restored by the addition of an extra Na'-ion. Proxy-aluminium is present in most of the complex amphiboles, for example, hornblende, as indicated in the formula quoted above.

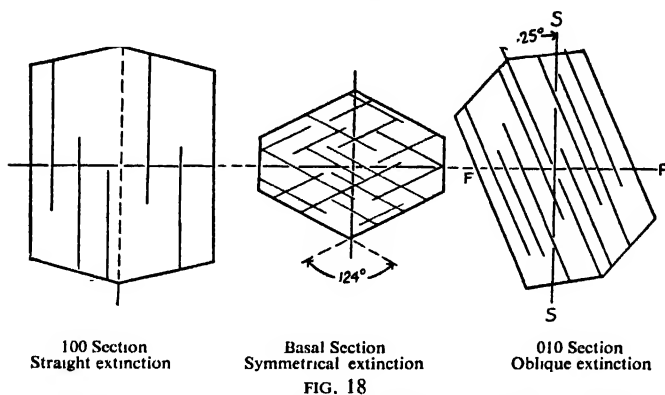
Further substitution of the same kind is seen in the formula of **pargasite**, an amphibole which is found in altered limestones, and which contains Al₂Si₆ instead of Si₈. The complete formula is $\text{NaCa}_2\text{Mg}_4\text{Al}(\text{Al}_2\text{Si}_6)\text{O}_{22}(\text{OH})_2$. An allied amphibole which occurs in alkali-rich igneous rocks is **hastingsite**. The chemical analysis of one specimen gave the formula: $(\text{Na},\text{K})_2\text{Ca}(\text{Mg},\text{Fe}''',\text{Fe}'',\text{Al},\text{Ti})_5(\text{Al},\text{Si})_8\text{O}_{22}(\text{OH})_2$. This is a formidable-looking formula which is included merely to illustrate the kind of ionic substitution which occurs in the amphiboles. Such substitution is controlled by two factors: the need to balance the total positive and negative charges, and the necessity for there to be sufficient spaces in the structure to accommodate all the ions. It will be noticed that in the last formulae there are eight instead of the usual seven cations. Actually in all amphiboles there are eight spaces which can be occupied by cations; but in some cases only seven of these are filled, one remaining vacant.

(2) Optical Orientation

In all but one rare type, the optical orientation of the amphiboles is the same as that of the pyroxenes: the optic axial plane lies parallel to (010). The details of extinction in variously orientated sections considered above in relation to the pyroxenes, apply equally to the amphiboles. The significant facts are illustrated in Fig. 18. The

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most obvious means of differentiating between the several members of the group is the size of the extinction angle in (010) sections, as



Sections through a crystal of hornblende, showing prismatic cleavage traces and drawn in positions of extinction. SS—slow vibration, FF—fast vibration.

indicated in Fig. 19. The angles read to the left of the middle line are those between the slow vibration direction (Z) and the cleavage traces which mark out the position of the vertical axis. Similarly,

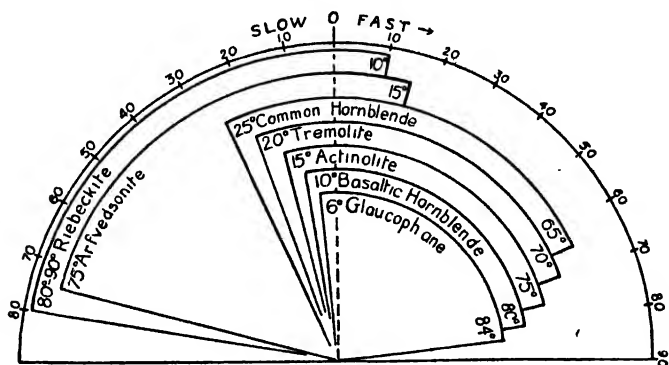


FIG. 19

Diagram showing extinction positions of the amphiboles. The figures in the left-hand top corner of each 90°-fan are the maximum extinction angles shown by (010) sections. The vertical line marked "O" represents the vertical cross-wire in the eye-piece of the microscope. The sides of the fans mark the positions of the cleavage traces in 010 sections in the extinction position.

those to the right, marked "fast" on the diagram, are measured between the fast vibration direction (X), and the cleavage traces. One significant point here is that while riebeckite and glaucophane

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both have much the same extinction angle, in the former the extinction position nearest to the cleavage traces is the fast vibration, but with the latter it is the slow direction.

Riebeckite.—Megascopically riebeckite closely resembles common hornblende, but under the microscope it is very distinctive. The colour is dull bluish green, and the absorption is so strong in some varieties that it appears nearly black. In less extreme cases the X vibration is deep indigo blue, Y is dull blue to brownish or

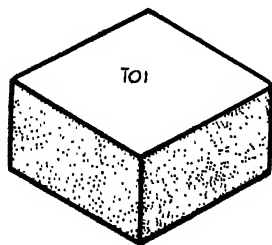


FIG. 20

The two ends of a simple crystal of hornblende, in plan. The forms represented are: clinoprism $\{011\}$ and hemi-orthodome $\{101\}$. The forms in the vertical zone are: prism $\{110\}$ and side (clino-) pinacoid $\{010\}$.

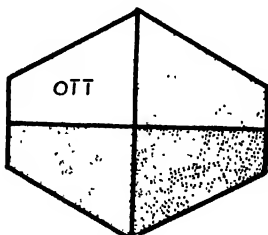
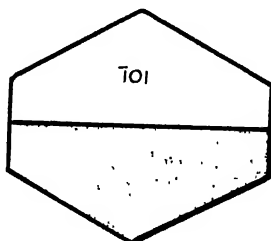


FIG. 21

The ends of a simple hornblende twin, in plan. The upper figure shows the hemi-orthodome faces; the lower, the clinoprism (or clinodome) faces. The twin-plane is $\{100\}$.

yellowish green in different cases, while Z is generally greenish, but may be nearly black. The birefringence is weak (0.004), so that in consequence of the strong absorption, there is little change of colour on crossing the nicols. Riebeckite is characteristic of strongly sodic acid igneous rocks: riebeckite-granite, microgranite and rhyolite have been described, the riebeckite-microgranites of Mynydd Mawr in North Wales (Fig. 84), and Ailsa Craig in the Firth of Clyde (Fig. 83) being good examples. In these rocks the riebeckite has a distinctive mode of occurrence: it forms micropoikilitic "mossy" patches in which the other components are embedded.

Riebeckite occurs on a much more important scale in many of the ring-complexes in northern and western Africa. A wide variety

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of riebeckite-bearing igneous rocks include "plutonic" riebeckite-granites, medium-grained dyke-rocks of like composition and extrusive riebeckite-rhyolites and pyroclasts. Many of these rocks are ideally fresh and provide admirable sections for petrological study. Finally riebeckite crystals up to several inches in length occur in associated pegmatites.

Glaucophane does not occur in igneous rocks, but is of special significance in the study of low-temperature, high pressure regional metamorphism which results in rocks of the glaucophane-schist facies. The rocks now represented by these schists were in part sodic igneous rocks, including spilites in the examples occurring in Anglesey. Optically glaucophane is most distinctive by reason of its striking pleochroism. In its typical form it exhibits the following: X, colourless to light yellow; Y, violet and Z, pure Prussian blue.

Barkevikite has been recorded from a number of British and foreign alkali-rich rocks, for example those occurring in the Lugar sill in Ayrshire. Again the pleochroism scheme is very distinctive, with X clear light yellow, Y strong red-brown and Z dark brown. The absorption formula is on the usual hornblende plan, with X the least, and Z the most strongly absorbed. Care has to be exercised in distinguishing between barkevikite and lamprobolite, commonly called basaltic hornblende. They are closely similar as regards pleochroism, absorption, extinction, position of the optic axial plane parallel to (010), and refractive indices; but differ in birefringence, basaltic hornblende (0.068) being much stronger than barkevikite (0.021) in this respect. There is need here for a lot more information.

Lamprobolite (basaltic hornblende): see preceding paragraph. This amphibole, which is brown in thin section, has been recorded from many different types of rock, but is dependent for its characters not so much on the kind of magma from which it crystallized, as on the conditions to which it was subjected subsequently. This is suggested by the fact that common hornblende can be converted experimentally into lamprobolite by heating, which has the effect of changing the state of oxidation of the iron, and of breaking down the hydroxyl, so that the $(OH)_2$ of the general formula becomes O'' .

Arfvedsonite is related chemically to barkevikite and occurs frequently in strongly sodic syenitic rocks such as sodalite-syenite and nepheline-syenite. It is very deeply coloured in thin section, a rich dark green, but differs from most other amphiboles in that the Z-vibration is the least absorbed. Further, arfvedsonite has negative elongation, the fast vibration, X, lying 15° or less from the cleavage traces.

Hastingsite is yet another alkali-rich aluminous amphibole occurring typically in nepheline-syenites (Ontario), theralites (Crazy Mts.,

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Mont.), etc. In its optical characters it approaches closely to pargasite and even common hornblende, but sections tend to be rich bluish-green in the position of maximum absorption. The extinction angle is a few degrees larger than that of hornblende; but the most distinctive feature is the very small optic axial angle. Thus, given a suitably orientated section, the identification of hastingsite, suggested by its associates, is readily confirmed.

(4) *Conditions of Formation and Distribution in Rocks*

Because of their complex composition and of the practical difficulties that arise, the conditions of crystallization of the amphiboles have not been studied experimentally in the same detail as the olivines and pyroxenes.

From consideration of their chemical composition it might be inferred that amphiboles would require a high flux-concentration for their development. The study of many thin sections containing both amphiboles and pyroxenes shows quite clearly that the former, represented by hornblende, follow pyroxenes, represented by augite, in the normal crystallization sequence. The amphibole may be moulded upon the pyroxene, or may show a replacive relationship towards it. There is, however, a third possibility: hornblende may be precipitated instead of pyroxene at temperatures of up to about 1,000° C. under plutonic conditions in the presence of adequate water. If the compositions of hornblendes are plotted together with those of pyroxenes on a CaMgFe triangular diagram, the former are seen to lie approximately midway between augites and hypersthene. Thus hornblende compositions bridge the immiscibility gap present in the two-pyroxene field described above. This means that a single hornblende may be regarded as the chemical equivalent of the two kinds of pyroxene.

The close chemical and structural similarity between the two groups results in particularly easy replacement of another kind. Under conditions of low-grade metamorphism pyroxene may be altered into an amphibole of appropriate composition, and usually of fibrous habit. This is the first stage in the conversion of a pyroxene-bearing basic rock into a hornblende-schist or amphibolite.

The reverse relationship may also be observed. Large euhedral hornblendes occurring in andesitic lavas may show all stages of conversion into granular pyroxene due to a temporary rise in the temperature of the lava, and release of pressure.

Common hornblende is characteristic of plutonic rocks, especially the acid and intermediate granodiorites, diorites and syenites. In the finer grained and more quickly cooled equivalents of these same rocks, the mineral assemblage has normally crystallized at a higher temperature, so that amphiboles are less common and may be

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absent. In alkali-rich igneous rocks such as nepheline-syenites, phonolites, etc., the special iron-rich, sodic types including primarily riebeckite, and less commonly hastingsite, barkevikite, arfvedsonite and kataphorite are widespread. Amphiboles are uncommon in basaltic and gabbroic rocks except those of alkaline facies.

MICAS

The micas constitute one of the most important groups of mineral silicates which are chemically distinct from the groups so far

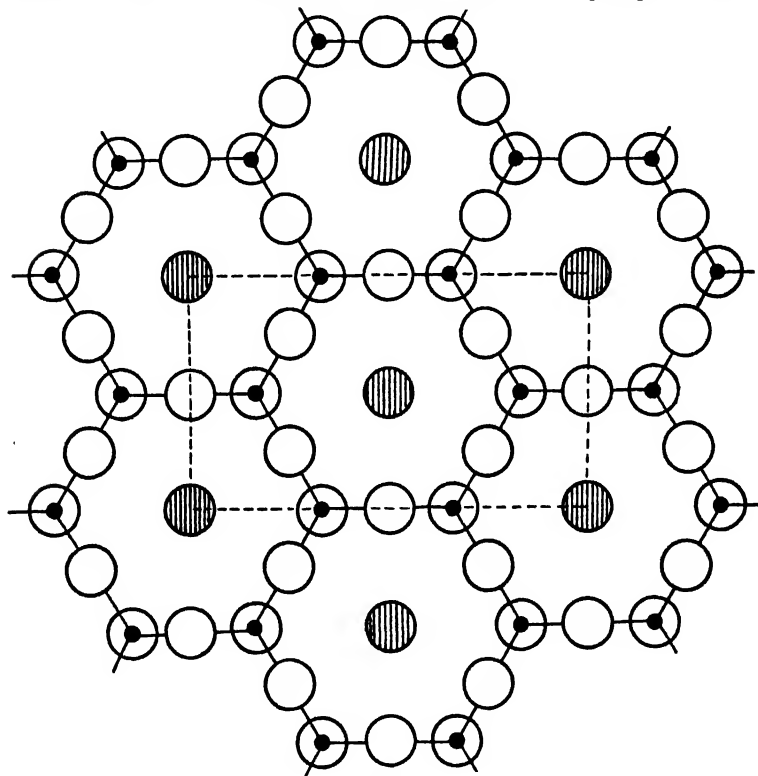


FIG. 22

Diagram showing the extended sheet structure of the micas.

Si atoms, black; O^{2-} atoms, white; hydroxyl (OH), ruled. The characteristic double sheet—the thinnest possible cleavage flake—comprises two of those illustrated, in mirror-image relationship (see Fig. 23).

considered in two respects: the alkali elements are important in all micas; but unlike certain amphiboles and pyroxenes, calcium is absent. In atomic structure, too, they are distinctive. The atoms are arranged in extended sheets as discussed below.

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Atomic Structure and Chemical Relations

In micas and in several other groups of minerals built on the same plan, the SiO_4 -tetrahedra are linked to one another at three corners, and thus form indefinitely extended sheets in which the atoms of different kinds are arranged on a hexagonal plan (Fig. 22). The sheets are actually double as shown in Fig. 23, and are "staggered"

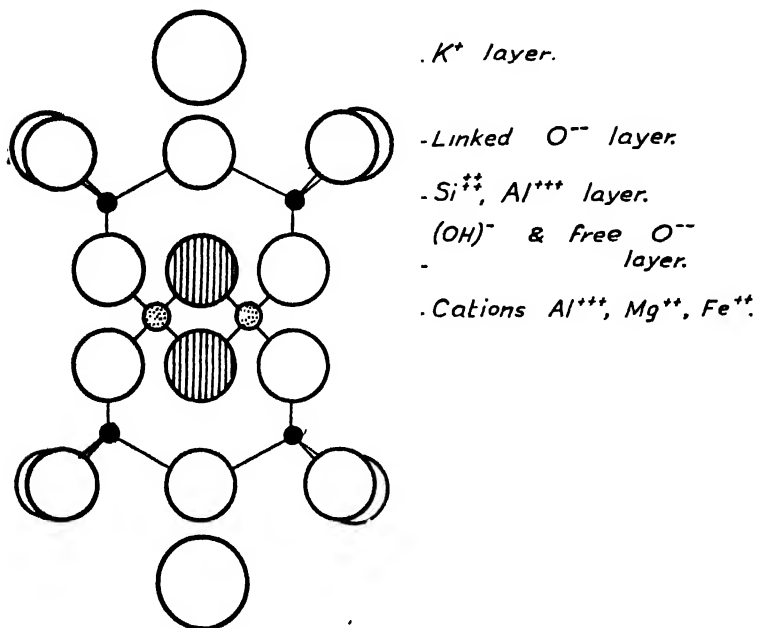


FIG. 23

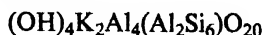
Diagram showing the order of atomic layers in mica: section through the double sheet.

relative to one another, which gives the crystals Monoclinic symmetry, although they are strongly pseudohexagonal. The unit of pattern contains Si_4O_{10} , but as with the amphiboles, to avoid having to deal with half atoms, it is convenient to double this. The essential hydroxyl groups are included in the planes containing the "free" oxygens in the tetrahedra, and four of these are associated with each unit.

In micas a certain degree of substitution of silicon by aluminium is normal. These Al^{+++} atoms which function as silicons are "proxy Al's." Each unit contains two of these Al-ions. Thus the "standard formula" is $(\text{OH})_4(\text{Al}_2\text{Si}_6)\text{O}_{20}$. There is a considerable residual negative charge on this unit which is balanced out by the addition of atoms of sodium and potassium, together with Al, Mg, Fe and

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Li in different cases. In all of the common micas, K' is the dominant alkali, occupying a unique position in the structure (Fig. 23). In muscovite, the remainder of the negative charge is balanced by Al, and the complete formula is:



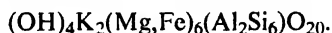
Actually a slight substitution of Al by Fe takes place in some muscovites, in which case $(\text{Al}, \text{Fe}'')_4$ may be substituted for Al_4 in the formula. Lepidolite is the best known "lithia-mica," but although repeatedly analysed, it has proved exceedingly difficult to express the composition in a simple formula. The following is an approximation:



with Li varying from one to four atoms, with corresponding adjustments in the Al, Si ratio.

Zinnwaldite is another lithia-mica, similar in general composition to the above, but with Fe'' replacing some of the Al.

If in the formula of muscovite Mg_6 replaces Al_4 , the formula of phlogopite is obtained: $(\text{OH})_4\text{K}_2\text{Mg}_6(\text{Al}_2\text{Si}_6)\text{O}_{20}$. This is often spoken of as the "magnesian mica." By comparison muscovite is the "aluminous mica," not the "potash mica," as it is so commonly called, for it contains neither more nor less potash than phlogopite or, indeed, than biotite. The latter is the common ferro-magnesian mica, and is of very complex composition. A first approximation to its composition is suggested by the following:



General Properties

As a direct consequence of their atomic structure the micas show pseudo-hexagonal symmetry, though actually they are monoclinic. To the eye they appear to be simple hexagonal crystals of tabular habit; but careful measurement shows that the basal pinacoid is not perfectly at right angles to the prism faces. The angle beta is within a few minutes of 90° however; and similarly the angles between the apparent hexagonal prism faces are nearly, but not exactly, 60° . Occasionally crystals are found which show hemibipyramid and other faces which betray the true symmetry of the mineral: otherwise the percussion figure obtained by smartly tapping a centre-punch placed in contact with the (001) face of the mica, proves that the symmetry is not hexagonal. A six-rayed star is produced in this way, but the rays are not identical: two, which lie in the single plane of symmetry, are more strongly developed than the others. This fulfils another useful purpose, as it serves as a guide when finding the position of the optic axial plane.

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The most striking physical property of all micas is the perfect basal cleavage, which takes place between the pairs of sheets of SiO_4 -tetrahedra, and in the planes of alkali atoms (Fig. 23). Not only are the atoms fewest in these planes, but the K' ions are very loosely bonded to the rest of the structure: they are in twelve-fold co-ordination.

Twinning is not uncommon in micas, on the so-called "mica-law," in which the twin plane is approximately coincident with (110), while the composition plane may be either (001) or (110). In the latter case the basal cleavages of both parts of the twin are co-incident, but on account of the different orientation of the two parts, there is a striking difference in absorption, best seen when looking obliquely through a cleavage plate towards a good light. Twinning is seldom seen in thin sections.

Optical Orientation

Although the pseudohexagonal symmetry of the mica crystal has been stressed above, examination of its optical properties immediately demonstrates that it is not Hexagonal. As a group the micas

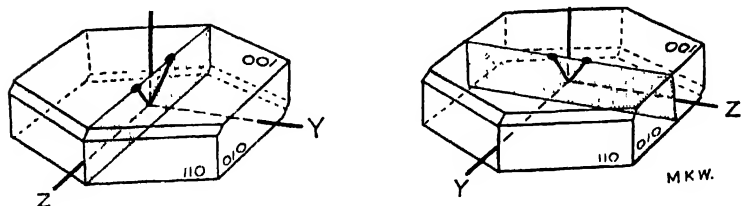


FIG. 24

Diagram showing the optical orientation of the micas. *Left*, the biotite-phlogopite series; *right*, the light micas.

are biaxial, though the size of $2V$ is very variable, and the optic axial plane in some cases is parallel, but in others perpendicular, to (010) (see Fig. 24). This difference in orientation is the basis of the division of the micas into two groups, which essentially comprise the light and the dark micas respectively. In all micas the fast vibration X is perpendicular to the (001) plane, and as X is the acute bisectrix, all micas are optically negative.

THE LIGHT MICAS

(a) **Muscovite** is the silvery white mica seen in hand specimens of many granites, particularly the strongly potassic varieties, and in

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the pegmatites and minor intrusives associated with them. In igneous rocks it is restricted to the most highly silicated types. It is also widely distributed in schists and gneisses.

Muscovite originates in another way, however. The potassic feldspars contain the same elements as muscovite, and are readily converted into the latter as a consequence of hydrolysis. Conversely, when rocks or even sediments containing white mica suffer thermal metamorphism, the mica is converted into orthoclase. A distinction is drawn between primary white mica formed by crystallization from a melt, and the secondary white mica produced by the alteration of alkali-rich silicates. The latter type is **sericite**. Although identical in composition with muscovite, sericite has an entirely different mode of occurrence in rocks, and there appear also to be slight physical differences. Sericite usually occurs in the form of aggregates of minute flakes. Soda-rich feldspars including albite may be crowded with such mica flakes, and this is believed to be a distinct species, known as **paragonite**, analogous in composition with muscovite but with Na substituted for K in the formula. There is some doubt about this substitution on account of the marked disparity in size between the K' and Na' ions. A recent analysis carried out on paragonite from paragonite-schist was shown to contain both K' and Na' ions, in the ratio of one to three. Paragonite is uniaxial or nearly so, thus differing markedly from muscovite which is biaxial, with $2V$ about 40° .

Optically muscovite is distinctive chiefly by reason of its strong birefringence—0.036. This ensures that most vertical sections show brilliant interference colours; but the basal section polarizes in light grey, and as the distinctive perfect cleavage traces are absent, it may easily be misidentified by the unwary. Such sections exhibit a perfect biaxial interference figure, however, on which the sign is easily checked.

(b) **Lepidolite**, a lithium-bearing light mica containing about 5 per cent LiO_2 , is attractively coloured lilac in the mass, and although large crystals a foot or more across are obtained from certain complex pegmatites, a more usual mode of occurrence is in the form of aggregates of flakes or scales, as the name implies. Some particularly attractive mineral specimens consist of lepidolite acting as matrix to brightly coloured lithium tourmalines. Although the general optical orientation is the same as for muscovite, the angle $2V$ is very small, and in some varieties the mineral is sensibly uniaxial. Lepidolite is a characteristic constituent in some complex pegmatites, in close association with Li-bearing tourmalines.

(c) **Zinnwaldite** is also a lithium-bearing mica (about 3 per cent LiO_2) but is iron-rich, whereas lepidolite is iron-free. It also occurs in pegmatites and is important in greisens.

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THE DARK MICAS

These stand in strong contrast to the group considered above, not only in appearance and optical orientation, but also in mode of occurrence and distribution in rocks.

(a) **Biotite**.—This is the common black mica that extends through the whole range of rocks from acid to ultrabasic. Biotite (or

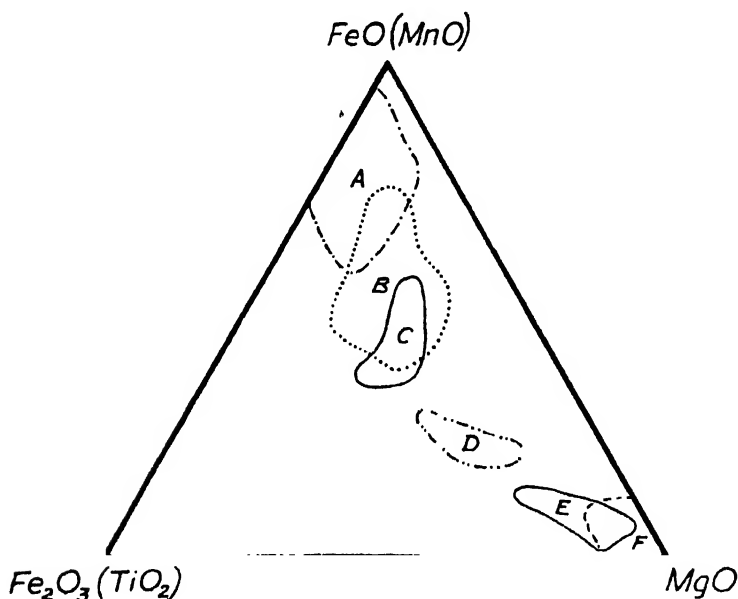


FIG. 25

Diagram showing the variation in composition of micas of the biotite-phlogopite series in rocks of different kinds.

- A—Dark micas from granite-pegmatites.
- B—Dark micas from granites, granodiorites, etc.
- C—Dark micas from tonalites and diorites.
- D—Dark micas from gabbros.
- E—Dark micas from ultrabasic rocks.
- F—Phlogopites from metamorphosed magnesian limestones.

Data by E. Wm. Heinrich.¹

glimmerite) consists almost solely of this mineral, while at the other end of the scale the black glistening plates of biotite immediately catch the eye in hand-specimens of granites, while some Indian pegmatites yield biotite crystals a yard across. Among many other biotite-rich rocks certain types of lamprophyres are outstanding. In these the crystals are often euhedral and zoning may be well

¹ "Studies in the Mica Group," *Amer. Journ. Sci.*, **244** (1946), p. 836.

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developed. Biotite frequently shows better than any other mineral the effects of internal bombardment by alpha particles emanating from radioactive elements in zircon inclusions. This bombardment causes concentric zones of discoloration round the inclusions, and is seen in thin sections as pleochroic haloes. Biotite alters rather readily into a light green chlorite, first along the cleavages, but ultimately the whole crystal may be replaced.

The optical orientation of biotite is as shown in Fig. 24. The angle $2V$ is very small, so that on rotation of the stage the isogyres must be closely watched to see that they do actually separate. Further, very strong illumination must be employed, as the absorption is particularly strong in basal sections, so much so that even in thin sections no light is allowed to pass in some cases. Vertical sections are almost violently pleochroic, the vibration parallel to X yielding a light (straw-) yellow absorption tint, while the Y and Z vibrations are usually indistinguishable, giving equally dark brown, sometimes nearly black. It follows that a basal section is non-pleochroic: there is no perceptible change on rotating such a section over the polarizer.

X-ray measurements suggest that high- and low-temperature forms of biotite occur, the former in igneous, and the latter in metamorphic rocks; but the fact is of no significance to the student, for the distinction can be made only by the specialist.

(b) **Phlogopite**, often called the bronze mica on account of its colour, is much less widespread in igneous rocks than biotite, and is in fact more characteristic of certain types of metamorphosed magnesian limestone, which in the hand specimen may be studded with small, perfect, bronze-coloured phlogopites set in a matrix of crystalline calcite. In igneous rocks true phlogopite is relatively rare, but occurs in mica-peridotites and in certain of the ultrabasic lavas in the Central African volcanic region and elsewhere. Thus in the micas as in the olivines and pyroxenes the highly magnesian types are found in the most basic rocks (Fig. 25).

CHAPTER TWO

THE FELSIC MINERALS

THE term felsic is mnemonic in the sense that it covers the minerals in the feldspar and feldspathoid groups, together with the silica group. For ease of reference the anhydrous aluminosilicates occurring in rocks may be listed in a manner which displays their chemical relationships, as under:

<i>Potassic</i>	<i>Sodic</i>
KAlSiO_4 .. kalsilite	NaAlSiO_4 .. nepheline
KAlSi_2O_6 .. leucite	$\text{NaAlSi}_2\text{O}_6$.. jadeite ¹
KAlSi_3O_8 .. orthoclase, etc.	$\text{NaAlSi}_3\text{O}_8$.. albite, etc.

It will be noted that the addition of silica (SiO_2) to kalsilite and to nepheline gives the second mineral in each column; and similarly the addition of silica to leucite and jadeite gives orthoclase and albite respectively. The SiO_4 and Si_2O_6 minerals in the list are unstable in the presence of free silica and are therefore said to be unsaturated. By contrast orthoclase and albite are stable under these conditions and are described as silica-saturated.

On account of their outstanding importance as rock-formers the feldspars are considered first, followed by the feldspathoids.

FELDSPARS

The feldspars are quantitatively the most important of the rock-forming silicates. They are all aluminosilicates—of potassium, sodium, calcium and rarely barium, and crystallize in the Triclinic and Monoclinic Systems. Although there must be differences in the details of the atomic structure of the Triclinic and Monoclinic feldspars, fundamentally they are much alike. The three-dimensional framework of linked SiO_4 tetrahedra may be resolved into “chains” of a special type, which are joined to one another by sharing oxygens² (Figs. 26, 27). The chains are aligned parallel to the *a*-axis

¹ Jadeite is included to show its chemical equivalence to leucite among the potassic minerals: it is actually related to the pyroxenes.

² Although we use the word “chain” for ease of interpretation, this is not comparable with the chains and bands of pyroxenes and amphiboles respectively, which were self-contained and joined to one another through the medium of the cations. In the feldspar structure, however, the “chains” are linked together, left and right, in front and behind, by sharing oxygens.

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(clino-axis) of the crystal, and two of these, in mirror-image relationship across a plane parallel to (010), occupy the unit cell.

Each single link of the chain consists of four SiO_4 -tetrahedra; but as all the oxygens are shared, it may be said to contain Si_4O_8 . Further, in each of these units, one Si is displaced by a proxy-Al, so the unit formula becomes $(\text{AlSi}_3)\text{O}_8$. This leaves a surplus negative charge, which in one important group of feldspars is balanced out by the addition of a K-atom, giving the formula of orthoclase: KAlSi_3O_8 . In another important member Na-atoms occur in the

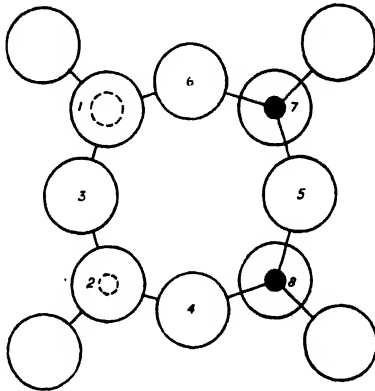


FIG. 26

A single link in the feldspar "chain": with eight oxygen atoms lying in the plane of the paper, two (Nos. 1 and 2) above, and two (Nos. 7 and 8) below the plane. Three silicons occur in the link (one hidden by oxygen 2), and one proxy-aluminium (behind oxygen 1).

place of K, giving albite, $\text{NaAlSi}_3\text{O}_8$. A third member is anorthite, which contains Ca in place of K and Na; but as Ca is divalent, an adjustment has to be made. A second proxy-Al''' is introduced in place of another Si''', so the complete formula becomes $\text{CaAl}_2\text{Si}_2\text{O}_8$. Finally, in rare cases Ba'' fills the role of Ca'', giving celsian, $\text{BaAl}_2\text{Si}_2\text{O}_8$.

Now as the K and Ba atoms are of approximately the same radius, they are capable of mutual replacement, and feldspars such as hyalophane occur, having a composition somewhere between orthoclase and celsian. Similarly, Ca'' can replace Na' in the structure to almost any extent, giving a number of minerals intermediate in composition between albite and anorthite, known collectively as the plagioclase series.

By contrast, Na cannot replace K in a manner which will retain the original structure, and therefore homogeneous Na-K feldspar

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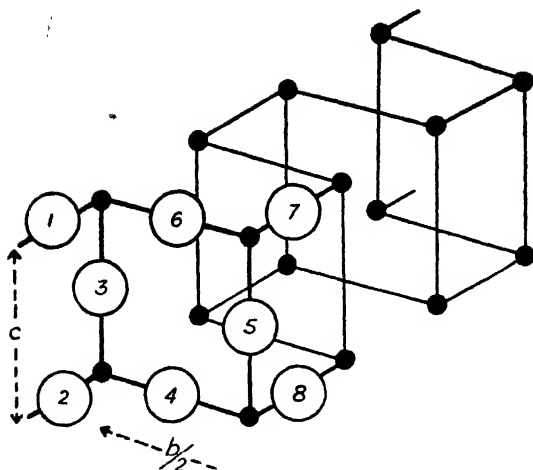


FIG. 27

A formal diagram of the "chain" structure of feldspar. Silicon and proxy-aluminium shown in black. To avoid overcrowding the diagram, only the "shared" oxygens in the single link, represented in Fig. 26, are included. The chain is half the width of the unit cell, and runs from back to front, parallel to the a -axis.

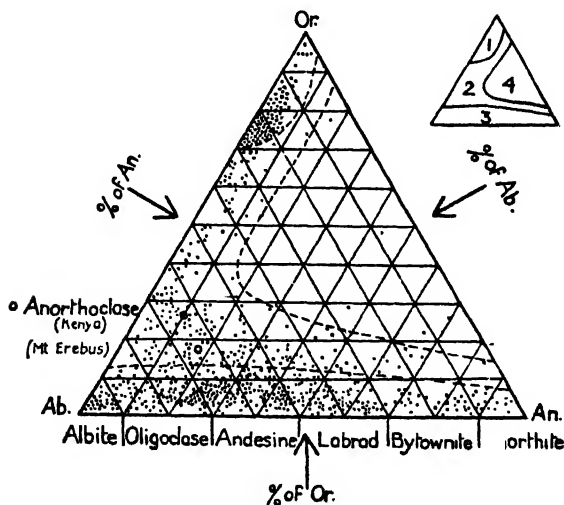


FIG. 28

Diagram illustrating composition of natural feldspars in terms of orthoclase, albite and anorthite. (After H. L. Alling and A. N. Winchell.) Note maximum concentration in fields 1 (orthoclase and microcline) and 3 (plagioclases).

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comparable with plagioclase does not occur, except at high temperatures (see below).

The formulae so far considered are those of "pure" feldspars, of ideal composition, and in the account which follows are represented by Or, Ab and An. Natural feldspars are actually ternary systems, consisting of various proportions of Or, Ab and An, and may therefore be represented by points on a triangle which has these three components at its apices. The three components are measured in the directions of the arrows in Fig. 28. The dots on the triangle represent the compositions of several hundred feldspars. The maximum concentration near the Or apex indicates the range of composition of the natural K-feldspars. The average K-feldspar can be seen to have the composition approximately of $\text{Or}_{76}\text{Ab}_{20}\text{An}_4$. The marked concentration of points just above the base-line indicates the range of composition of the plagioclases. It will be noted that on average the plagioclases, although fundamentally Na-Ca feldspars, contain about 5 per cent (or more) of Or. The absence of points in field 4 (Fig. 28) reflects the fact that Ca'' and K' , on account of dissimilarity in atomic radius, are not mutually replaceable in the atomic structure.

General Characters of the Feldspars

In the Monoclinic feldspars the simplest combination of faces consists of the basal pinacoid (001), side-pinacoid (010) and prism (110). A slightly more complicated crystal is illustrated in Fig. 29.

In the Triclinic feldspars although crystal faces are developed in analogous positions, and although they superficially resemble the Monoclinic crystals, the lower symmetry causes the Monoclinic prism to be represented by two complementary hemi-prisms, "m" and "M." Similarly the hemibipyramid of the Monoclinic feldspars becomes the quarter-bipyramid of the Triclinic types. The better cleavage in all feldspars is parallel to (001); but that parallel to (010) is little inferior. Naturally these two cleavages are at right angles in Monoclinic feldspars, but intersect at angles between approximately 93° and 94° in the plagioclases.

THE ALKALI FELDSPARS

In this category are included the potassic (K') feldspars, the sodic (Na') feldspars and the several intergrowths between them, termed, generically, perthites.

(1) The Potassic Feldspars

Several distinctive species or polymorphs of potassic feldspar exist, all having the same essential composition, KAlSi_3O_8 , but with slight differences in the positions of the ions in the atomic structures,

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which are determined mainly by the temperature of crystallization. Sanidine, orthoclase and adularia are Monoclinic, while microcline is just Triclinic, as the name implies.

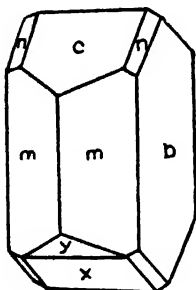


FIG. 29

Crystal of orthoclase (Carlsbad habit).

c Basal pinacoid (001).

b Clinopinacoid (010).

m Prism (110).

x and *y* Hemi-orthodomes (20 $\bar{1}$) and (10 $\bar{1}$).

o Hemi-bipyramid (1 $\bar{1}$), in zone with *x* and *b*.

n A clinodome (or clinoprism) (011).

(a) **Orthoclase.**—This form of KAlSi_3O_8 occurs in a variety of crystal habits. The two commonest are the so-called Carlsbad habit (Fig. 29) in which the crystals are tabular, flattened parallel to the side-pinacoid faces (010); and the Baveno habit in which the crystals

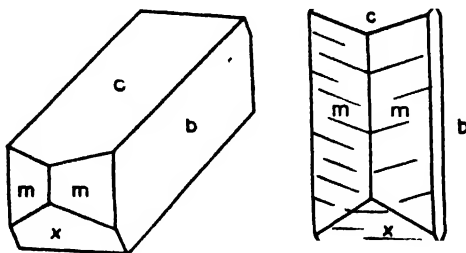


FIG. 30

Left, Orthoclase of Baveno habit. Right, Adularia.

are elongated parallel to the inclined *a*-axis. A third habit is characteristic of the feldspar adularia, and crystals of this type are prismatic, while the side-pinacoid faces are virtually suppressed. These crystals have a roof-like termination formed by the basal pinacoid and the hemi-orthodome, dipping in opposite directions: they can be easily identified provided the cleavage traces are visible (see Fig. 30).

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Orthoclase crystals are frequently twinned in distinctive fashion (Fig. 31), on one or other of the following laws:

Carlsbad, usually interpenetrant and resulting from rotation about the *c*-axis;

Baveno, in which twin-plane and composition-plane are parallel to a clinodome (011);

Manebach, in which twin-plane and composition-plane are parallel to (001).

Of these the first is much the commonest; Manebach and Baveno twinning are uncommon in orthoclase, but less rare in sanidine.

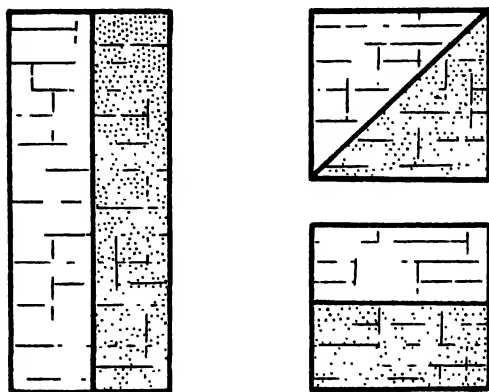


FIG. 31

Vertical sections through twins of orthoclase.

Left, Carlsbad; Right top, Baveno; Right bottom, Manebach.

Optical Characters

The refractive indices of orthoclase are low: γ , 1.525; α , 1.519, both well below Canada balsam. Birefringence also is weak (0.007)—a little lower than the quartz by which it is normally accompanied in igneous rocks. The optic axial plane is perpendicular to the plane of crystallographic symmetry (010), and is inclined at 5° to 8° to (001). Therefore the extinction is oblique to this extent in (010) sections. The angle $2V$ is large; but, on heating, it progressively diminishes, and becomes 0° at a certain temperature, above which the axes open out in the symmetry plane (010) (see Fig. 32).

(b) **Sanidine.** On heating, orthoclase inverts into sanidine at the inversion temperature, 900°C . As might be expected, therefore, sanidine is the form of K-feldspar¹ occurring in quickly-cooled lavas

¹ The high temperature structure of sanidine allows Na and K ions to occupy equivalent positions, so that, although most sanidines are dominantly potassic, they may contain appreciable Na.

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and some dyke-rocks, including rhyolites, trachytes, pitchstones, etc. It may persist indefinitely in these rocks, in the metastable state; but with the slow cooling characteristic of deep-seated rocks sanidine inverts into orthoclase. The only significant difference between the two minerals in thin section is the difference in orientation of the optic axial plane, noted above, and the small, to very small, $2V$ in sanidine. When unweathered, sanidine is clear and glassy.)

(c) **Adularia**.—This is the name of a crystal habit rather than a mineral species, though it has been found that adularia is a variety of microcline (rather than orthoclase) containing about 10 per cent

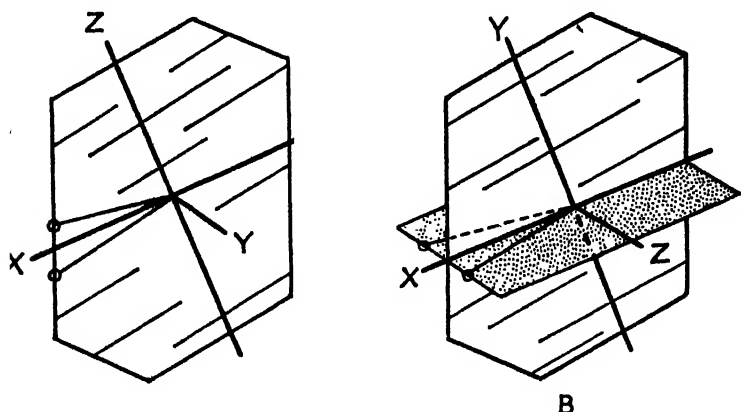


FIG. 32

Diagram showing optical orientation of sanidine (A), and orthoclase (B).
Optic axial plane stippled in B.

of albite. It is of prismatic habit (Fig. 30) and commonly occurs in mineral veins and as incrustations in vugs. It may be glassy like sanidine or opaque like orthoclase.

(d) **Microcline** is the Triclinic form of K'-feldspar, but the departure from Monoclinic symmetry is very slight, the angle between the basal and side pinacoids being $89^{\circ} 50'$, so that there is no visible difference between orthoclase and microcline in the hand specimen, even if a contact goniometer is available for measuring.

Under the microscope, however, microcline is very distinctive by reason of a complex system of spindle-shaped twin lamellae in two sets nearly at right angles in (001) sections (Fig. 33). The twinning may affect only part of a crystal, or it may be absent altogether, but when found, it is completely diagnostic. Refractive indices (γ 1.529, α 1.522) and birefringence (0.007) are very similar to those of orthoclase, but on account of the Triclinic symmetry, the extinction

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in a (001) section is oblique (extinction angle 15° to the (010) cleavage traces).

Although most types of microcline are coloured red or brown, specimens from pegmatites from Colorado and elsewhere are bright green in the variety known as **amazonstone**. This contains the rare elements caesium and rubidium, and the depth of colour is said to be proportional to the amount of the latter.

Microcline is practically restricted to the highly potassic granites

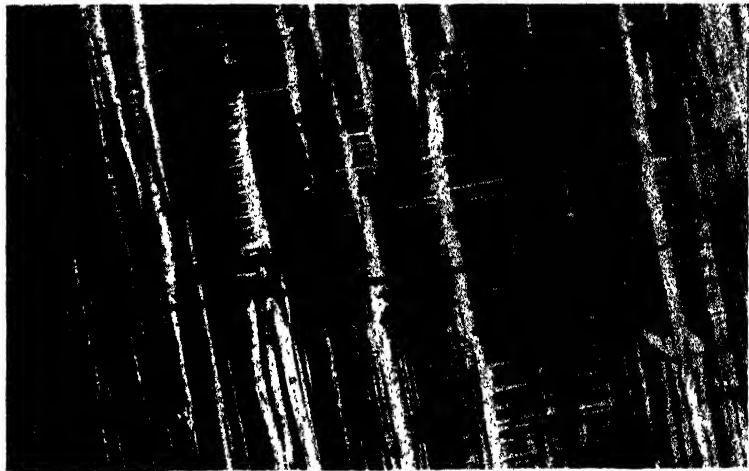


FIG. 33

Thin section of microcline-albite perthite, Hylba, Ontario.
The original is a typical green amazonstone, with megascopic veins of white albite.

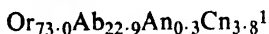
Crossed polarizers. Photomic. by E. O. Rowland. Section parallel to (001).

and pegmatites among igneous rocks; but it occurs in many metamorphic rocks, often in large porphyroblastic crystals formed by potassium metasomatism. In granites it is rarely pure microcline but contains the substance of albite intergrown as perthite, while even in those specimens which appear homogeneous, analysis often proves some substitution of Na' for K'.

The extent to which the barium feldspar, **celsian** (Cn), $\text{BaAl}_2\text{Si}_2\text{O}_8$, enters into the composition of potassic feldspars is not known with certainty because in many analyses barium is not separately determined. It is known, however, that certain types of feldspathoidal lavas are relatively barium-rich, and the barium is presumably in

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the alkali-feldspar. For example "sanidine" occurring in a rock of this type has the following composition:



(2) *The Sodic Feldspars*

As in the case of the potassic feldspars, there are several structural states in which $\text{NaAlSi}_3\text{O}_8$ can occur, dependent upon the conditions of crystallization. The differences between the varieties are very slight, however, and not nearly so apparent as those, for example, between orthoclase and microcline. Two varieties of different habit may be recognized in albite and clelandite: the former is of outstanding importance.

(a) **Albite.**—The status of albite is somewhat anomalous: it is the end-member of the important plagioclase series and is therefore indubitably a plagioclase; but in composition it is analogous with orthoclase, with which it is commonly intergrown in varying proportions. Natural albite invariably contains some orthoclase and conversely. Therefore it is important to regard albite as an alkali feldspar rather than as a plagioclase.

Albite is stable under widely varying conditions. It occurs as a primary constituent in many rock-types, particularly in alkali-granites and syenites: indeed one variety of the latter consists of nothing but albite and is therefore termed albitite. Secondary albite is also widely distributed in rocks of a wider range of compositions which have suffered the type of change referred to as albitization. Albite is also an important constituent of some varieties of crystalline schists. In its crystallographic and other physical properties albite is like any other plagioclase, as described below.

(b) **Clelandite.**—This mineral bears the same relationship to albite as adularia bears to orthoclase: it occurs chiefly in pegmatites and sometimes in mineral veins, in aggregates of tablet-like crystals flattened parallel to (010). Again it is the name of a crystal habit rather than of a distinct mineral species.

(3) *The Sodipotassic Feldspars: the Perthites*

As noted above both orthoclase and microcline are liable to form intergrowths with albite. The latter occurs in the form of films and, in the coarser types, less regular plates or veins. The films are usually of constant orientation dependent upon the structure of the host mineral (Figs. 34 and 35). The following types of perthite have been described and figured: vein-perthite (probably the commonest), film-perthite, and braid-perthite in which the albite sheets are

¹ Anal. Scoon, J. H., quoted from Tilley, C. E., *Amer. Min.*, **43** (1958), pp 758–61.

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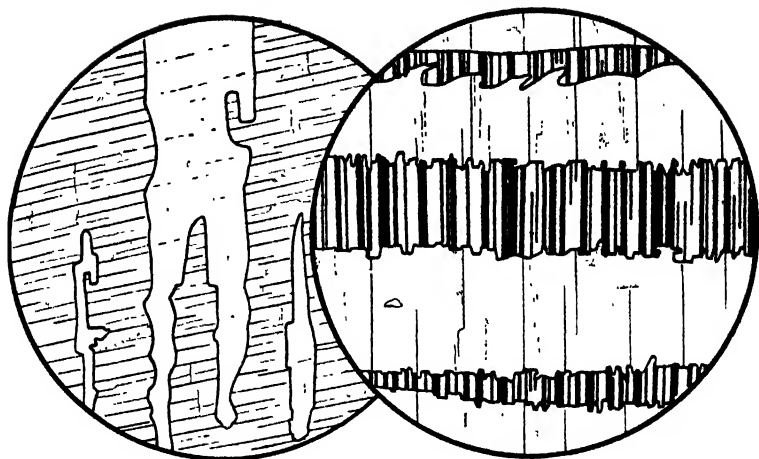


FIG. 34

Formalized sections through vein perthite (albite veins in microcline). *Left*, a (010) section in plane polarized light; (001) cleavage shown. *Right*, a (001) section showing the nature of the margins of the albite veins against the host-microcline, and the (010) cleavage-traces.

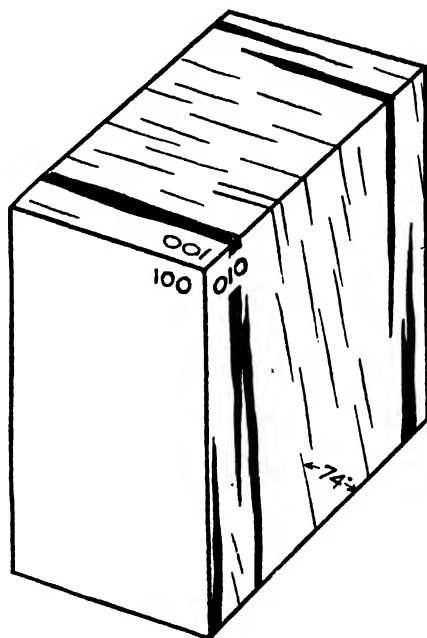


FIG. 35

Block-diagram showing relation of film- to vein-perthite in K-Na feldspar.

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arranged parallel to the prism faces (110) and (110), and thus give a braided appearance in a basal section. Less regular intercalations are known as patch- and string-perthite which, on account of the lack of orientation, may have originated differently. The intergrowths may be megascopic, microscopic, or on so fine a scale as to be beyond the resolving power of the microscope, though "visible" to X-rays. The terms perthite, micropertthite, cryptopertthite and X-ray perthite are applied to these, respectively; and in the case of microcline being the host feldspar, the mineral name is added as a qualifier: thus, "microcline-micropertthite" indicates an intergrowth of microcline and albite on a microscopic scale, the albite lamellae varying from 5 to 100 microns in thickness. If albite is the host mineral to inclusions of potassic feldspar, the term antipertthite is used. The significance and mode of origin of perthitic feldspars are discussed below under the heading "Stability Relationships."

The terms soda-orthoclase and soda-microcline are sometimes given to cryptopertthites of the appropriate composition: they are chemically alike.

A striking optical effect is produced when the intergrowth is on a very small scale (cryptopertthitic). Thus *moonstone*,¹ a semi-precious gemstone, is prized for its bluish sheen which results from the submicroscopic interlamination of two feldspars with slightly different refractive indices: the effect is called schillerization.

THE PLAGIOCLASE SERIES

It has long been known that the members of this series show a gradational variation in composition between the two end-members, albite and anorthite. For descriptive purposes it is convenient to use names to fix certain ranges of composition. Using the convention already described, we may note that, ignoring the small amount of Or present, the composition of any specific plagioclase may conveniently be expressed by a simple symbol, indicating the molecular percentages of Ab and An which it contains. Thus the full range extends from $Ab_{100}An_0$ to Ab_0An_{100} . The several kinds of plagioclase are defined arbitrarily by erecting divisions at Ab_{10} , 30, 50, 70 and 90. These are the widely agreed limits between albite, oligoclase, andesine, labradorite, bytownite and anorthite respectively. The percentage composition of three selected plagioclases is tabulated below:

	SiO ₂	Al ₂ O ₃	Na ₂ O	CaO
Albite (Ab)	68.7	19.5	11.8	0.0
Labradorite ($Ab_{50}An_{50}$)	55.6	28.3	5.7	10.4
Anorthite (An)	43.2	36.7	0.0	20.1

¹ Spencer, E. Edmonson, ". . . moonstone from Ceylon . . ." *Min. Mag.*, 22 (1930), p. 291.

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Since the ratio of the molecular weight of *Ab* to *An* is as 1 : 1.061, it is approximately accurate to calculate the relative weights of the two components (*Ab* and *An*) in the proportion of the number of molecules. Thus $\text{Ab}_{60}\text{An}_{40}$ contains *Ab* three-fifths and *An* two-fifths by weight. These partial analyses stress the fact that Na'-rich plagioclases are Si-rich, but Al-poor, while Ca''-rich types are correspondingly poorer in Si but richer in Al.

Formerly it was believed that the plagioclases provided the perfect example of an isomorphous series, involving progressive replacement of Na'Si'''' (100 per cent in albite) by Ca''Al'''' (100 per cent in anorthite). This has proved to be an over-generalization: only parts of the full compositional range are isomorphous. Two breaks occur, at approximately An_{30} and An_{70} . Plagioclases less calcic than An_{30} are isomorphous except for a limited range of composition, between An_5 and An_{17} , which is a zone of unmixing occupied by the so-called peristerites—beautifully schillerized albite-oligoclase feldspars. Similarly plagioclases more calcic than An_{70} (that is, those covered by the terms bytownite and anorthite) are isomorphous. Between the two breaks, and therefore covering the range andesine-labradorite, the plagioclases are heterogeneous: they consist of *two* feldspars having the composition and structure of albite and anorthite respectively. The interlamination of the two is submicroscopic, but can be inferred from X-ray analysis.

But in spite of these breaks the overall effect is to produce a series of feldspars which, for all practical purposes, are isomorphous; so that the optical properties with which the student is mainly concerned vary progressively and systematically from one end of the series to the other (see p. 79).

Crystallographically the plagioclases differ from orthoclase by having the basal pinacoid (and the cleavage parallel to it) inclined to the side-pinacoid (010) at an angle of 86° to 87° instead of 90° ; otherwise they are closely similar in appearance and cleavage. Plagioclase may be twinned in any of the ways in which orthoclase twins, but in addition exhibits two special types which are distinctive. Plagioclase of any composition may be twinned once or repeatedly on the Albite Law, in which twin plane and composition plane are (010). Secondly, in twinning according to the Pericline Law, twin plane and composition plane vary systematically in the manner shown in Fig. 36. The pericline, like the albite twinning, is commonly repeated, and in both cases the twin-bands are thin, and give a characteristic lamination between crossed polarizers (Fig. 37). Both types may be associated with simple twinning on the Carlsbad Law (Fig. 37). Sometimes these features can be seen with the naked eye, but more readily with a lens, while they make the plagioclases unmistakably distinctive in thin sections.

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Statistical studies of the frequency of occurrence of the several types of twins shows that normally twinning is more complex in igneous than in metamorphic rocks; in those of high grade, as compared with low-grade metamorphism; and in plagioclase of calcic,

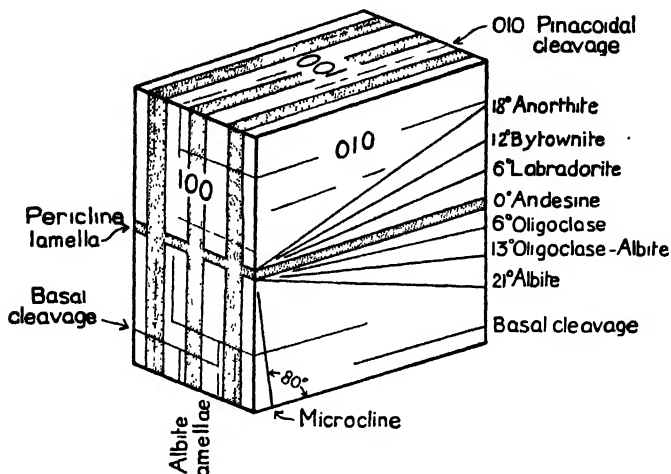


FIG. 36

Block-diagram showing relation between albite and pericline twinning and the cleavages in plagioclase feldspars.



FIG. 37

Twinned crystal of plagioclase feldspar in andesite from Cupar, Fife.
(Photo by Geol. Survey.)

rather than sodic composition. Thus, in low-grade metamorphic rocks untwinned grains are often dominant, and may be exclusive—a condition that is not found in igneous rocks. At a slightly higher grade of metamorphism, simple twinning is characteristic, but is on

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the albite, NOT the Carlsbad Law. In high-grade metamorphic rocks, the complexity may approach that of igneous rocks, but still there are differences. Complex twinning involving combinations of simple (Carlsbad and/or Manebach) and lamellar (albite and/or pericline) twinning is commonly encountered in sections of igneous rocks particularly those of basic composition; but this feature is generally absent from rocks of metamorphic origin.¹

Plagioclase crystals often contain minute inclusions which may be arranged irregularly, or in definite planes parallel to the periphery of the crystal. The latter is commonly the case with phenocrystic plagioclase in basalts, when the inclusions may be glass, devitrified glass or small patches of the groundmass. In deep-seated rocks, such as certain gabbros and norites, the plagioclase may be strongly schillerized and exhibits a striking play of colour in the hand-specimen. In thin slice this is seen to be due to myriads of orientated rod-like inclusions, apparently of iron-ore. It seems probable that the iron was in solution in the feldspar at high temperatures (proxying for Al⁺⁺⁺). On heating, the schillerization due to these inclusions disappears and the iron is reincorporated in the feldspar, which is converted into a high temperature form.

A definitely zonal structure is more commonly shown by plagioclases than by any other type of mineral.² This results from variation in chemical composition during crystal growth, and may be of several different kinds: the commonest involves a gradation from a Ca-rich core to Na-rich outer layers, and is termed *normal zoning*. Less frequently the converse is true: this is *reverse zoning*; while in yet other cases the composition has changed rhythmically, causing *rhythmic* or *oscillatory zoning*. The systematic study of such zoning in plagioclase may throw considerable light on the details of the cooling history of the rock which contains it. Studies of this kind, and indeed normal petrographical work, involves the accurate identification of the plagioclase, or in a zoned crystal, of its individual layers.

Plagioclase is liable to alteration, and occasionally suffers complete replacement by secondary white mica, scapolite, zeolites, or minerals of the epidote group, often associated with calcite. The last type of alteration results from dynamothermal metamorphism, and results in the separation, as it were, of the Ab from the An. The former is stable; but the latter changes readily under these conditions into zoisite, clinozoisite or epidote, embedded in a "background" of

¹ Turner, F. J., "Twinning of Plagioclase in Metamorphic Rocks," *Amer. Min.*, **36** (1951), p. 581, with refs.

² Phemister, J., "Zoning in Plagioclase Feldspar," *Min. Mag.*, **23** (1934), p. 541; Fries, C., "Resorbed Feldspars in Basalt," *Amer. Min.*, **24** (1939), p. 782; Hills, E. S., "Reverse and Oscillatory Zoning in Plagioclase Feldspars," *Geol. Mag.*, 1936, p. 49.

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secondary albite. The name *saussurite* is sometimes applied to plagioclase showing such alteration.

Status and Distribution of Plagioclase in Rocks

Plagioclase of one kind or another occurs in representative members of all the main rock groups. Rarely it may make up nearly the whole of the rock: albitite, oligoclase, andesine and labradorite (under the name anorthosite)-have all been described, though only the last is other than very rare.

Apart from these monomineralic types, however, plagioclase is an important—often a dominant—component in many intermediate and basic rock-types. In general, in passing from more acid to more basic types, the plagioclase becomes progressively richer in An: thus in syenitic rocks oligoclase is commonly found; in dioritic (including andesitic) types, it is andesine typically; while in gabbroic (including noritic and basaltic) rocks, labradorite or bytownite occur. Anorthite is less common: indeed, though it does occur in some gabbroic rocks (such as allivite, for example), it is more typical of metamorphic rocks. An argillaceous limestone might be expected under thermal metamorphism to give rise to anorthite, among other Ca-rich minerals.

Identification of Feldspars in Thin Section

The identification of feldspars in rock sections involves techniques which must be mastered by students at an early stage of their training as such identification lies at the root of rock classification and naming.

It should be realized at the outset that some grains may be so orientated as to make summary identification impossible. Twinning is often sufficiently diagnostic, particularly the distinctive “cross-hatching” of microcline and the lamellar twinning of plagioclase; but not all microcline, and not all plagioclase is twinned in this way. Further, any kind of feldspar may be untwinned; or may be lying with the twin-plane parallel to the plane of the slide, in which case the crystal is apparently, though deceptively, simple. Such “simple” crystals must be viewed with suspicion: they may be orthoclase, sanidine, microcline or plagioclase, and before they are identified as, say, untwinned orthoclase, a confirmatory test must be made. The Becke Test is most useful in this connection, and is quickly carried out, preferably on grains lying on the edge of the slide, so that direct comparison may be made between the refractive indices of the grain and of the embedding medium.¹ The indices of

¹ Remember that the bright line moves into the substance of *lower* refractive index on racking *down*, with the light well diaphragmed down to ensure maximum definition.

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orthoclase and microcline are well below that of the embedding medium; the table in Fig. 38 shows the range of R.I.s of the plagioclases.

Actual measurement of the R.I.s of feldspars in a rock must be carried out on grains separated from a crushed specimen. Careful

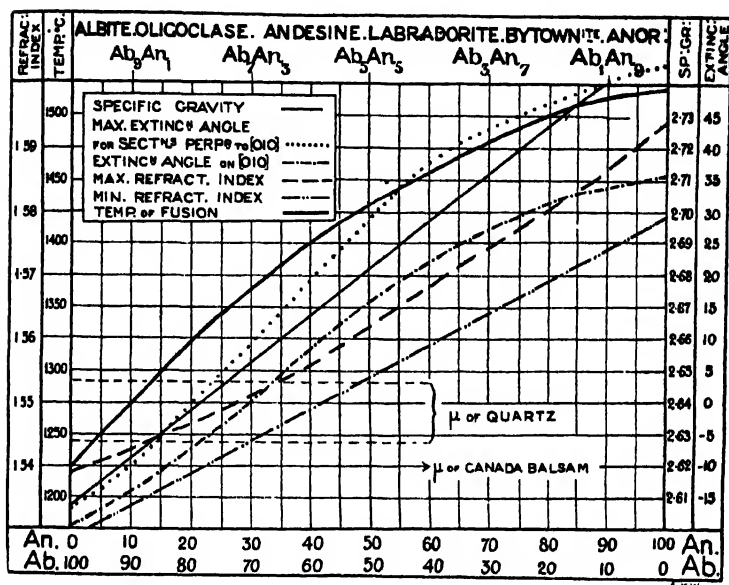


FIG. 38

The physical characters of the plagioclase feldspars.

(Modified after H. L. Alling.)

measurement gives results comparable with those obtained by chemical analysis.

If, on inspection of a slide, it is believed that two feldspars are present, it is useful to realize that it is virtually impossible for both to have the same "habit" and to be altered to the same extent and in the same way. Figs. 80 and 107 are convincing in this respect.

Given that plagioclase has been recognized in a rock, it remains to discover what kind of plagioclase before the rock can be accurately named. Restricting ourselves to optical methods using the ordinary petrological microscope, the composition of the plagioclase is indicated by the maximum extinction angles obtained from suitably

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orientated sections. A useful method makes use of sections at right angles to the Albite-twin lamellae. Such sections show equal illumination, and therefore the same shade of grey, when the twin planes lie parallel to the vertical cross-wire. They also extinguish symmetrically: the angles obtained on either side of the cross-wire should be approximately the same. Several measurements should be made on different grains, as only the *maximum* angle is diagnostic. Reference to the line on Fig. 38 marked "maximum extinction angles for sections perpendicular to (010)" will then give the composition of the plagioclase. In the special case of oligoclase ($Ab_{80}An_{20}$) the extinction of both sets of twin lamellae is straight. As the composition changes towards albite, the extinction angle increases to a maximum of 20° , while andesine ($Ab_{62}An_{38}$) also gives this angle. For angles less than 20° , therefore, a confirmatory test must be applied. This is easily done by means of the Becke Test; the refractive indices of andesine are above, while those of albite are below, Canada balsam.

Extinction angles measured on complex Carlsbad-Albite twins can also be used to determine composition: the method is fully described—together with several others—in all determinative mineralogy textbooks, to which the reader is referred.

Stability Relationships

The equivalence of sanidine (high-temperature K'-feldspar characteristic of quenched rocks) and orthoclase (low-temperature K'-feldspar of the same composition but occurring in coarse-grained and more slowly cooled igneous rocks) has long been known. Similarly it has been realized that the alternative crystallization of either orthoclase or microcline must be controlled by the physical conditions of temperature and pressure (doubtless by little understood kinetic factors of crystallization), as indicated by the fact that microcline is the typical K-feldspar of low-temperature granites, granite-pegmatites and a wide range of metamorphic schists and gneisses. Because temperature-controlled differences of atomic structure are much more subtle in their effects in the plagioclase series, it is only comparatively recently that they have been recognized. Differences were first suspected because of the lack of agreement in detail between the optical orientation of plagioclase phenocrysts from quenched lavas and those from coarser-grained plutonic rocks. These differences are greatest for sodic plagioclases; but even so they are so slight that they can be detected only by accurate Universal Stage measurements. X-ray investigation is needed to establish the structural state of the feldspars; and the complexity of the problem is indicated both by the enormous

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amount of research that has been devoted to it and by the fact that interpretation of much of the data is still in doubt.

Details of the changes affecting the feldspar structures subjected to variations of temperature are imperfectly known and largely of interest to specialists¹; but one of the main factors involved can be easily understood by reference to the simplified diagram of atomic structure, Fig. 26. This shows the cation positions occupied by Si^{++++} and Al^{+++} . In the diagram the latter is shown in the top left position; but it may occupy any of the four cation positions. In high-temperature feldspars the arrangement of the Al^{+++} ions is random, and the structure is described as "disordered." By contrast, in low-temperature feldspars the structure is "ordered"—the Al^{+++} ions occupy the same relative positions in all the (AlSi_3) units.

In the case of the end-members of the feldspar group, pure K-, pure Na- and pure Ca-feldspar, the possible structural changes are limited largely to the degree of ordering of the Al^{+++} ions. It is only in the case of the K-feldspars that the effects of this can be detected easily from the optical differences between K-sanidine and orthoclase, and by the optical and crystallographic differences between these two minerals and microcline. In this series orthoclase has a structure which is transitional between the disordered one of sanidine and the ordered one of microcline.² Orthoclase actually grades into microcline with an increase in the so-called triclinicity of the crystals as the ordering of the structure becomes more perfect. Generally the maximum triclinicity of the lowest temperature microcline is accompanied by the most perfect development of cross-hatched twinning: crystals which are transitional between orthoclase and microcline may have the twinning only patchily developed. At the highest temperatures, above $800^\circ\text{C}.$, the stable form of K-feldspar is "high sanidine," which is different in some details of optical orientation from "low sanidine," stable between $800^\circ\text{C}.$ and $650^\circ\text{C}.$, which is approximately the temperature at which sanidine inverts into orthoclase. The corresponding inversion temperature between high and low albite is $720^\circ\text{C}.$ ³

Discovery of transitional optical and crystallographic features is of great petrological significance: it indicates that the crystal

¹ Of the many papers on feldspar structure, see, e.g., Ferguson, R. B., Trail, R. J., and Taylor, W. H., "The crystal structures of low-temperature and high-temperature albites," *Acta. Cryst.*, 11 (1958), pp. 331-48.

² It has been suggested as an alternative that orthoclase results from sub-microscopic and repeated twinning of units of the Triclinic microcline structure, giving overall Monoclinic symmetry.

³ These figures are approximate only and are affected by pressure, flux concentration and the amount of K-feldspar in solid solution. They are included only to give the reader an idea of the temperatures involved.

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concerned was originally precipitated at a high temperature and that, due to the sluggishness of the transformation, relics of the high-temperature state have survived. A completely stable, low-temperature state is much less informative. This may arise either (a) because very slow cooling has allowed an originally high-temperature crystal to invert completely, or (b) because the original crystallization was at a temperature below the inversion level. There is no way of distinguishing between these alternatives. Exactly the same considerations apply to high- and low-albite, although, as noted above, the differences between the structural states in this case are so slight that they can be distinguished only by specialized tests.

When we come to consider the feldspars of mixed composition—the normal condition—the opportunities for differences to develop in the degree of ordering of the structures are greatly increased. Not only can Al''' occupy various positions as outlined above; but this also applies to the other cations. The effects are most marked in alkali feldspars. At high temperatures the Na' and K' occupy similar and interchangeable positions in the structure, which is therefore disordered. This gives a single type of homogeneous structure for the whole of the alkali feldspar (sanidine-anorthoclase) series. As the temperature falls, interchangeability of the alkali ions becomes restricted so that the Na' and K' are gradually forced to occupy different parts of the structure. Eventually this gives rise to unmixing (see below) and the growth of separate structural units of sodic and potassic feldspars, each in a more perfectly ordered state than previously.

Unmixed alkali feldspars are the **perthites** which have been described above. Perthites in which either orthoclase or microcline form the host mineral with inclusions of albite, are the slowly cooled and low-temperature equivalents of sanidine; while antiperthites with sodic plagioclase as host mineral are equivalent to anorthoclase.

The process of unmixing may start when the temperature falls below a critical value represented by the highest point on the solvus curve¹ (such as shown in Fig. 72 to which the reader may refer at this point); the first structures to form are sub-microscopic and detectable only by X-ray study. Whether or not further unmixing will occur depends upon the rate of cooling: if this is rapid it may enable sanidine to survive as a metastable mineral. The cooling rate is one of the most important factors in controlling the effectiveness of unmixing, and in general the coarsest intergrowths occur in very slowly cooled rocks.

¹ The position of this curve, representing the phase boundary limiting the ex-solved or unmixed phases (i.e. the limits of solid solution), is by no means certain. Heating natural perthites until they become homogeneous gives a very different maximum for the solvus temperature than that obtained with synthetic crystals.

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A series of perthitic structures of increasing coarseness can be recognized as follows:¹

X-ray perthites;

Cryptoperthites which give rise to distinctive optical effects such as the schillerization of moonstones; but the structure is too fine to be distinguished microscopically;

Microperthites, with intergrowths on a microscopic scale;

Perthites, megascopic intergrowths recognizable in hand-specimens.

The finest intergrowths (X-ray and cryptoperthites) can be produced experimentally by slowly cooling a homogeneous sanidine to below the unmixing temperatures, and conversely, such perthites can be homogenized on reheating. Unmixing of the coarser perthites is not reversible under experimental conditions. This is due primarily to the time factor: because of the very slow rates of ionic diffusion involved in the unmixing process, structures developed in natural perthites which have crystallized and cooled during periods of tens, hundreds or even thousands of years cannot be put into reverse by reheating for the duration of a laboratory experiment.

Tuttle has put forward convincing arguments for believing that the unmixing process does not stop at the formation of perthites, and he has suggested the addition of two further stages to the series listed above. In the first the Na ions migrate to the margins of the host crystals of K-feldspar, to form marginal zones or rims of albite. In the final stage, the completely unmixed albite and K-feldspars may recrystallize as adjacent discrete crystals. Thus it is possible that the combination of separate crystals of Na- and K-feldspars found in some granites may have developed by a long process of unmixing and recrystallization in the solid state from an original single phase of sodi-potassic feldspar.

This interesting hypothesis cannot be proved experimentally for the feldspars; but its truth is suggested by analogy with some sulphide systems in which the rates of unmixing are much faster. Mixtures of bornite and chalcopyrite heated to the melting point and then quenched at 600° C. form homogeneous solid solutions. Heating again to 600° C. and cooling under controlled conditions, ranging from five minutes to twenty-four hours, results in the development of a series of structural arrangements exactly analogous to those described above for the alkali-feldspars, including intergrowths of varying degrees of coarseness; and with more protracted cooling, rims of the minor component around cores of the more abundant

¹ Tuttle, O. F., "Origin of contrasting mineralogy of extrusive and plutonic salic rocks," *Journ. Geol.*, 60 (1952), pp. 107-52; also: Tuttle, O. F., and Bowen, N. L., "Origin of granite in the light of experimental studies in the system $\text{NaAlSi}_3\text{O}_8\text{-KAlSi}_3\text{O}_8\text{-SiO}_2\text{-H}_2\text{O}$," *Geol. Soc. Amer. Mem.*, 74 (1958).

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mineral; and finally aggregates of discrete grains of bornite and chalcopyrite in close association, but yielding no trace of having evolved from high-temperature solid solutions. Perhaps the final observation is the most significant for petrologists.¹

From this very brief survey it can be appreciated that the interpretation of the history of crystallization of a feldspar from its present state is beset with difficulties. This is emphasized by the fact that Bowen and Tuttle list fourteen² possible phase combinations that may be found in alkali feldspars. Even a single crystal may contain four recognizable phases, as for example the combination: high-albite, low-albite, orthoclase and microcline.

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In this group are included several minerals which are closely related, as their name implies, to the feldspars. They contain the same elements as the latter, though in different proportions, and are notably poorer in silica.

Leucite, the potassic feldspathoid, resembles orthoclase in composition. The SiO_4 -tetrahedra are linked corner to corner, and if we consider three of these, the unit contains Si_3O_6 . But in each unit of structure one proxy-Al takes the place of one Si; so that the formula of the unit becomes $(\text{AlSi}_2)\text{O}_6$. Balance of the total positive and negative charges is brought about by adding one K'-ion, giving the ideal formula, $\text{K}(\text{AlSi}_2)\text{O}_6$. In natural leucites some Na ions invariably replace some of the potassium, so the formula is more accurately represented as $(\text{K},\text{Na})\text{AlSi}_2\text{O}_6$.

It is difficult to write about the crystallography of leucite in unequivocal terms. In many mineral collections large crystals occur in the form, apparently, of the simple unmodified icositetrahedron; but the optical properties of the crystals prove that they cannot belong to the Cubic System: the form is a *pseudo-icositetrahedron*, resulting from complex twinning of Orthorhombic or possibly Monoclinic crystals which, in the untwinned condition, are unknown in Nature.

In thin slice leucite normally shows characteristic eight-sided sections, when cut through centrally. The refractive index, 1.508, is one of the lowest among common rock-forming minerals. Cleavage is absent; but small inclusions may be arranged in zones or tangentially, especially in the minute leucites occurring in the ground-mass of some lavas. The most significant feature, however, is the twinning (Fig. 39). In a perfect, centrally-cut section twin lamellae

¹ Schwartz, G. M., "Intergrowths of bornite and chalcopyrite," *Econ. Geol.*, 26 (1931), pp. 186-201.

² *op. cit.*, p. 27.

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in six directions (parallel to the faces of the pseudo-rhombdodecahedron) give a particularly striking appearance, though, on account of the abnormally weak double refraction (0.001), very strong illumination is necessary to make this feature really convincing. The mere fact that the twin planes are parallel to directions which, in a Cubic crystal, would be planes of symmetry, proves that such leucites are not Cubic.

As a fact, the small leucites in the groundmass of a lava do appear to be isotropic between crossed Nicols, and therefore are presumably

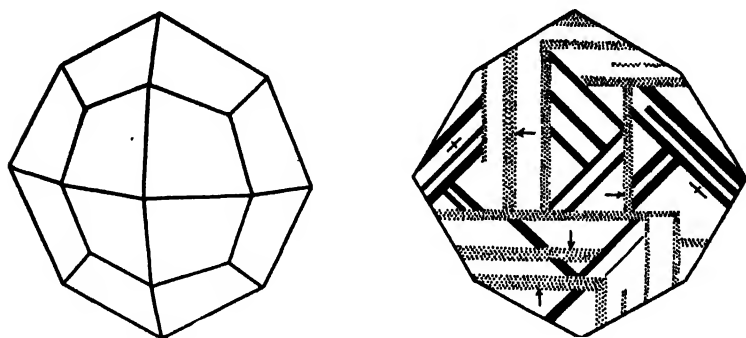


FIG. 39

Left, crystal of leucite showing pseudo-icositetrahedral form. *Right*, central section through a leucite crystal between crossed polarizers to show twinning in six directions. Dip arrows and signs for verticality indicate the attitude of the twin lamellae in relation to the section.

Cubic. This is notably the case with lavas in which the groundmass crystals are embedded in a base of glass. These facts seem to establish the existence of two forms of leucite: a high-temperature Cubic form, only seen in Nature's quenching experiments; and a low-temperature form occurring as phenocrysts, and exhibiting the twinning described above.

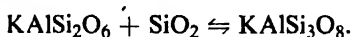
Remarkable pseudomorphs after leucite have been described from strongly potassic rocks in the medium and coarse grained categories from several different localities, one (doubtfully) British. In some instances the original icositetrahedral shapes have been perfectly preserved; but at the other extreme only a vague suggestion of the original form remains and poorly defined, rounded, light-coloured masses are *inferred* to be pseudoleucites from their mineral composition. These pseudomorphs consist of orthoclase or sanidine, nepheline and a small amount of adventitious material representing original impurities. In a recent description the composition was found to be: sanidine 66 per cent, nepheline 30 per cent and acmite

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3 per cent.¹ This composition is, of course, consistent with that of the original mineral, and is evidently an equal volume change resulting from complete reconstruction during slow cooling. We would emphasize that this change has nothing whatever to do with weathering: it represents an adjustment to changed environmental conditions and is, in effect, unmixing carried to the ultimate limit. Pseudoleucite represents the ultimate mineral association stable under the existing low-temperature, atmospheric pressure conditions.

These pseudoleucites demonstrate one of the most important facts regarding the stability of leucite: it cannot survive under conditions involving the retention of volatiles. As the concentration of the latter increases, the stability field of leucite diminishes, as shown in Fig. 72, and is ultimately completely eliminated, irrespective of the degree of undersaturation of the magma. If the latter is sufficiently undersaturated to yield *potential* leucite, the place of the latter must be taken by some other mineral or combination of minerals: biotite, by virtue of its composition, is one of those best suited to fill this role.

Leucite is thus virtually confined to rocks that have crystallized under low-pressure conditions, *i.e.* lavas of the appropriate composition which have been quenched. It is one of the first minerals to be precipitated from a magma of the right composition (*i.e.*, potash-rich) as indicated by its invariable occurrence as euhedral phenocrysts. If cooling is sufficiently slow, reaction with silica in the melt converts some or all of it into orthoclase. This is the reverse of the process known as incongruent melting of orthoclase which involves the reaction:



On heating orthoclase to approximately 1,170° C. it has been shown experimentally² to melt incongruently to leucite crystals in liquid which contains the excess SiO₂ liberated by the reaction. It is important to realize that at high temperatures leucite can exist in the presence of free silica. If at this stage quenching ensues, these leucites will survive as phenocrysts, while the liquid congeals as glass. With slow cooling, however, at the appropriate temperature the leucites will be made over into orthoclase—provided the magma contains sufficient silica. If it does not, some leucite will remain, in association with orthoclase. The relationship between leucite and orthoclase is thus analogous in all respects to that between olivine and pyroxene.

¹ Zies, E. G., and Chayes, F., "Pseudoleucites in tinguaitite from the Bearpaw Mts., Arkansas," *Journ. Petr.*, 1 (1960), pp. 86-98.

² Morey, G. W., and Bowen, N. L., "The melting of orthoclase," *Amer. Journ. Sci.* (1922), pp. 1-22.

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Summarily, therefore, there are three factors which control the formation and survival of leucite: (1) the degree of saturation of the potassic magma; (2) the rate of cooling; (3) water-vapour pressure. It may be noted that the occurrence of nepheline is dependent upon only the first of these factors: therefore it is much more widespread, in a greater variety of rocks in all grain-size groups, than leucite. Leucite is best developed in certain lava-types occurring in volcanic regions of the Roman province in Italy, the Leucite Hills in Wyoming, the Kimberley District in Australia and lava fields in Central and Eastern Africa. The mineral lends its name to two rock-types, leucitite (which ought to consist of leucite only, but does not) and leucitophyre. Certain Vesuvian lavas are well known by reason of the perfectly formed phenocrysts of leucite which they contain embedded in a dark basaltic-looking matrix: leucite-basanite is one such type.

Kalsilite is the only rock-forming mineral with a really helpful name: the first five letters indicate the chemical formula, which is ideally KAlSiO_4 , though actually considerable atomic substitution of Na' for K' occurs in all known natural kalsilites. This mineral was originally misidentified and described as nepheline, quite understandably, as it shows identical crystallographic characters and is optically indistinguishable: X-ray analysis was necessary before its identity was established.¹

It is known to occur in only four localities, three of them in Central Africa and one in Italy, and in all cases the mode of occurrence is the same: it forms complex phenocrysts closely associated with nepheline, leucite and sometimes sodalite in rare ultrabasic lavas.

Nepheline in the pure state may be represented by the formula NaAlSiO_4 , but it is isomorphous with kalsilite, and all natural nepheline contains some K' -ions, substituting for Na' . In phenocrysts from a recent Central African lava the proportions of the two components have been found to correspond with the formula $\text{K}_3\text{NaAl}_4\text{Si}_4\text{O}_{16}$. The same relationship occurs between nepheline and kalsilite as between albite and orthoclase: both pairs form mix-crystals at high temperatures, and with falling temperature unmixing occurs resulting in the formation of a complete series of intergrowths, from X-ray perthites, cryptoperthites, to microperthites. A further degree of unmixing gives rise to euhedral rimmed crystals, the core consisting of kalsilite and the rim of nepheline. Neither mineral is "pure": the core contains K_{80} , while both the rim and groundmass

¹ One cannot help wondering whether the mineral identified on the usual optical criteria as nepheline is really that mineral or the optically indistinguishable kalsilite. Apart from X-ray study, chemical analysis should show any discrepancy between the content of K_2O and the amount of nepheline (?) in the rock.

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nepheline contains between a quarter and a half of kalsilite. Using the same convention regarding composition as for the feldspars, the groundmass nephelines range from $\text{Ks}_{25}\text{Ne}_{75}$ to $\text{Ks}_{45}\text{Ne}_{55}$.¹

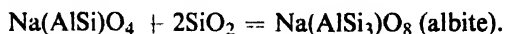
Diagnostic Characters of Nepheline

Nepheline crystals which the student is likely to encounter during his training are restricted to thin rock-sections. They are known to belong to one of the less symmetrical classes of the Hexagonal System; but in thin section they look like simple stumpy hexagonal prisms, about as broad as they are tall. Consequently square (vertical) and six-sided (basal) sections are typically well displayed. The former display first order (generally grey) interference colours, and of course extinguish straight; the latter are isotropic and give a negative uniaxial interference figure. The latter is one of the most useful characters which distinguish nepheline from untwinned feldspar for which it may quite easily be mistaken on account of the similar low relief and weak birefringence. Cleavage is variable. In some cases the square sections show cleavage traces, the hexagonal ones do not: therefore the cleavage is pinacoidal, parallel to the basal plane (0001). In other cases comparison of the vertical and basal sections demonstrates clearly that the cleavage is prismatic. Usually the cleavage traces are very feebly developed, but they may be made clearer by alteration. This may occur in a variety of ways. Nepheline may be replaced by an aggregate of white mica flakes or by zeolites or cancrinite. The latter is very distinctive and when it rims nepheline is a very useful aid in confirming the diagnosis of nepheline in doubtful cases.

As might be expected nepheline crystals are frequently zoned, though it needs expert handling of the microscope to make the zoning visible. It can be rendered much more striking if the nepheline is stained with fuchsine after treatment of the rock-section with weak acid.²

The Role of Nepheline in Igneous Rocks

Nepheline occurs in rocks characterized by an abundance of alkalies and alumina, but low in silica. Thus it is never found in Acid rocks, but may be abundant in those of Intermediate to Basic composition. In an acid magma containing potential free silica combination takes place between the components of nepheline and the silica, thus:



¹ Sahama, Th. G., "Kalsilite in the lavas of Mt. Nyiragongo, Belgian Congo," *Journ. Petr.*, **1** (1960), p. 146.

² Shand, S. J., "Staining of Feldspathoids and on Zonal Structure in Nepheline," *Amer. Min.*, **24** (1939), p. 508.

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Consequently nepheline tends to take the place of albite in magmas deficient in silica, and in certain types of nepheline-syenites, for example, nepheline may be the dominant felsic mineral. In hand-specimens of nepheline-syenites and nepheline-gabbros the mineral may be more obvious even than in thin slice, particularly on weathered surfaces. In certain examples from near Pretoria, South Africa, the rock-surfaces have been naturally etched; the white feldspars and black coloured silicates stand in strong relief to the euhedral nephelines up to half an inch across, which have been weathered-in, but show their characteristic shapes. In fresh specimens the nephelines are grey, with a rather silky texture and greasy lustre; but weathering develops a strong red or reddish-brown colour.

Among corresponding lavas, a well-characterized type is phonolite (Fig. 99); while among the Basic lavas nepheline occurs in tephrites and basanites. The term nephelinite is applied to similar lavas in which the nepheline is unaccompanied by feldspar of any kind.

Cancrinite is related to nepheline in composition, and in a sense may be regarded as "nepheline-carbonate" with a rather doubtful formula, perhaps $3\text{NaAlSiO}_4 \cdot (\text{Ca}, \text{Na}, \text{H})\text{CO}_3$.

This mineral is Hexagonal, but rarely forms definite crystals. Occasionally it is a primary constituent of nepheline-syenites, and in so-called cancrinite-syenites is an essential component. More often it fills a minor role, as an alteration product of nepheline, or arises by reaction between the latter and included grains of calcite (Fig. 97). In Britain cancrinite has been recorded from a pegmatitic facies of borolanite, at Loch Borolan, Assynt.¹

Under the microscope it occurs usually in shapeless masses, either in, or marginal to, nepheline. It is colourless, with both refractive indices below balsam: $\alpha = 1.496$, $\gamma = 1.519$. It has a perfect prismatic cleavage, and as its birefringence is high (0.023) it may resemble muscovite, but is easily distinguished therefrom by its negative relief, by its associates, and by showing a uniaxial negative interference figure in conveniently orientated sections in convergent light.

Three other members of the Feldspathoid group of minerals are often associated with nepheline and are closely related in chemical composition. They are sodalite, nosean and hauyne.

Sodalite.—Sodalite is a Cubic mineral which is normally seen as grains or interstitial patches in certain types of nepheline-syenites. Occasionally it ranks as an essential mineral and then may be seen to form bright blue patches in the hand-specimen.

¹ Stewart, F., *Min. Mag.*, 26 (1941), p. 1.

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In thin section sodalite is invariably colourless, and has a particularly low refractive index (1.48). This is considerably lower than the indices of the feldspars and nepheline with which the sodalite is normally associated. An imperfect dodecahedral cleavage may be visible in some sections. Between crossed polarizers the mineral is, of course, isotropic.

On account of its rather negative characters, sodalite is never easy to identify with certainty, and in some cases a micro-chemical test is necessary. Small quantities of the mineral have been discovered in certain rocks by using ultraviolet light, when sodalite exhibits a brilliant yellow fluorescence.

The exact composition is rather doubtful: it may be represented by the following formula:



Nosean or **Noselite** is also Cubic, and crystallizes in the same form as sodalite, *i.e.*, the rhombdodecahedron. The crystals are commonly very light yellowish-grey in thin section, and although they may show the characteristic six-sided sections to be expected on account of the crystal form, they are often strongly corroded, with a heavy dark margin (Fig. 40). As the refractive index is low (1.490) negative

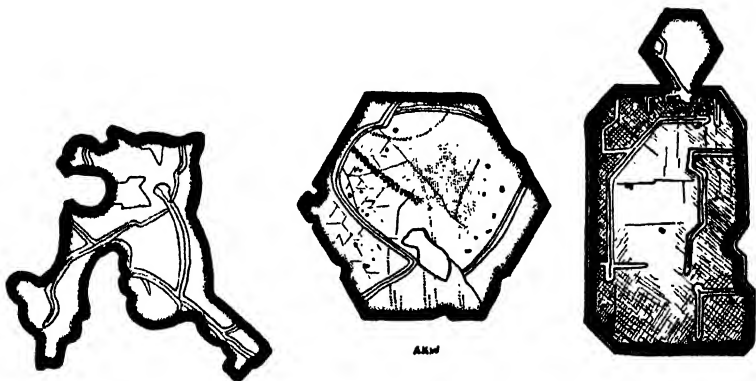


FIG. 40

Sections of nosean from nosean-leucitophyre, Rieden, Eifel, showing effects of magmatic corrosion, the characteristic canal-like markings, trains of gas bubbles, dodecahedral cleavage and (*right*) schiller structure.

surface relief is exhibited. Curious canal-like channels cross the surface, and traces of a dodecahedral cleavage can frequently be discerned.

Hauyne or **Hauynite**.—This mineral also crystallizes in the same form as sodalite and nosean—the rhombdodecahedron—and shows

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much the same features as the latter, in regard to shape, corrosion, and cleavage, but is distinctive on account of its sky-blue colour in thin section.

As regards composition, again there is some doubt, but both are closely related to nepheline, and while nosean contains Na_2SO_4 , hauyne contains CaSO_4 , in both cases additional to the nepheline molecule. Actually natural specimens are intermediate between these two, owing to the interchangeability of Na and Ca.

Both nosean and hauyne are essentially restricted to fine-grained nephelinitic or leucitic rocks. The former occurs, for example, in the nosean-leucitophyre from Rieden in the Eifel (Fig. 100), and in the nosean-phonolite which forms the Wolf Rock off the Cornish coast. Hauyne is not uncommon among the leucitic lavas of the Roman volcanic province.

Melilite.—This name is applied to a series of uncommon, somewhat complex silicates of calcium, aluminium and magnesium. Any one specimen may be regarded as having a composition that can be expressed in terms of the two end-members of the series, akermanite ($\text{Ca}_2\text{MgSi}_2\text{O}_7$) and gehlenite ($\text{Ca}_2\text{Al}_2\text{SiO}_7$).

It will be realized that, although gehlenite is closely comparable with anorthite and to that extent might be grouped with the feldspathoids, melilite of average composition differs from all the other feldspathoids by containing Mg⁺⁺.

Melilite crystallizes in one of the less symmetrical classes of the Tetragonal System, but crystals are rare, and the student is likely to encounter the mineral only in thin sections. It is exceedingly rare in igneous rocks and is virtually restricted to certain ultrabasic lavas and dyke rocks: only one named rock-type in the coarse-grained category bears melilite—it glories in the name uncomphagrite. Many of the known occurrences of melilite-bearing rocks are found in the great lava fields in central and eastern Africa, where the mineral is associated with leucite, kalsilite, nepheline and the accessory, perovskite—all unsaturated minerals, incapable of existing in the presence of free silica. An interesting British occurrence of melilite has been described from the contact zone of a Tertiary basaltic plug intrusive into Chalk at Scawt Hill near Belfast, Ireland.¹

In thin section melilite is very distinctive. The sections are roughly rectangular with rather irregular margins; the relief is moderately high, but the chief diagnostic feature is the anomalous birefringence. The typical interference colour is a deep inky blue, irregularly distributed: the marginal parts tend to be lighter than the central zone which may be nearly isotropic. There is no other mineral with which

¹ Tilley, C. E., "The Dolerite-Chalk contact of Scawt Hill . . .," *Min. Mag.*, 22 (1931), p. 439.

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melilite may be confused. Its constant companion is **perovskite**, (CaTiO_3), occurring in small octahedra, with exceptionally dark borders due to its very high refractive index.

The genesis of melilite has been studied experimentally and is discussed in the petrographical section of the book.

THE SILICA GROUP

Several distinct mineral species consist of pure silica, SiO_2 ; some of these occur as good crystals, others as cryptocrystalline aggregates, while amorphous silica also occurs naturally.

Each of the three mineral species, quartz, tridymite and cristobalite, occurs in both high- and low-temperature modifications, distinguished as beta (or high-) and alpha (or low-) quartz, etc. The complete range of six minerals forms a series stable under varying physical conditions.

In all these silica minerals SiO_4 -tetrahedra are linked to one another by all their corners. The actual arrangement of the atoms is complicated and impossible to illustrate simply. The important fact is that tetrahedra are linked spirally in both forms of quartz, the arrangement being somewhat more symmetrical in the high- than in the low-temperature form: thus while α -quartz crystallizes in the holo-axial (trapezohedral) class of the Trigonal System, β -quartz belongs to the corresponding class of the Hexagonal System.

Alpha-quartz is the quartz of mineral veins and vugs: it is usually a product of inversion in igneous rocks. Its crystal characters are indicated in Fig. 41. Usually there is a distinct difference in the degree of development of the two rhombohedra (positive and negative); and special forms, particularly trigonal trapezohedra and trigonal bipyramids, show these crystals to be fundamentally different from those of β -quartz.¹ The latter shows a characteristic Hexagonal bipyramidal form with sometimes a poorly developed Hexagonal prism in addition: these are forms characteristic of the holosymmetric class of the Hexagonal System; but actually it is known from the internal structure of the mineral that it belongs to the trapezohedral class of that system.

Twinning is apparently ubiquitous in α -quartz, according to a variety of laws; but it can only be demonstrated by means of external characters in some cases—notably by the occurrence of the form (51 $\bar{6}$ 1) on contiguous faces, instead of on alternate ones (Fig. 41C).

¹ Quartzes from the iron mines of north-western England show the simple crystallographic character normally associated with the β -phase, but they are definitely low-quartz. Apparently the presence of the haematite affects the crystal habit of the silica.

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Otherwise the twinning can be demonstrated by special optical tests¹—not by the examination of sections of normal thickness.²

The study of twinning in β -quartz has been somewhat neglected as the phenomenon produces no visible effects in thin rock-sections: but it has been shown to be quite normal. Collections of milky

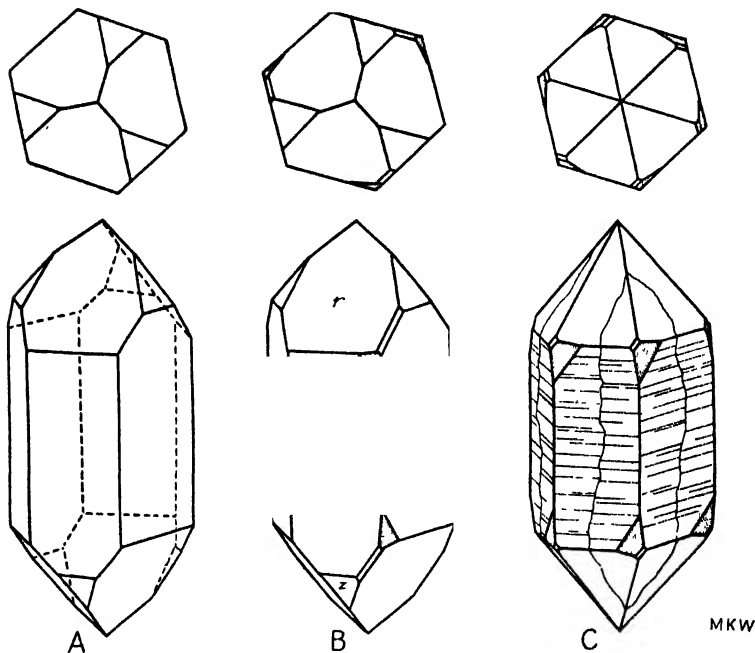


FIG. 41

Quartz crystals: (A) Simple combination of positive (r) and negative (z) rhombohedra and vertical prism (m); (B) A right-handed crystal showing the same forms, with trapezohedron ($51\bar{6}1$) and trigonal bipyramid; (C) A double left-handed Dauphiné twin. *Trigonal trapezohedron faces stippled.* (After Parrish and Gordon.)

crystals of β -quartz from Cornish quartz-porphyrries contain many twins, the twin-plane being either a face of a rhombohedron or a trapezohedron.³ The angle between the c -axes of the two crystals involved varies in different types between 43° and 84° . The so-called Japanese twins are of these types.

The inversion of quartz from the "high" (β) to the "low" (α)

¹ By immersing a thick basal section of quartz in oil of the same refractive index and viewing it along the direction of the optic axis between crossed plates of polaroid.

² For much interesting information on the twinning, etc., of quartz, see "Symposium on Quartz Oscillator Plates," *Amer. Min.*, 30 (1945), p. 205.

³ Drugman, J., "On β -quartz twins from some Cornish localities," *Min. Mag.*, 21 (1927), p. 336.

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form is achieved by only a slight distortion of the atomic structure, and so occurs readily at a fixed temperature which may be used as a point on a geological thermometer. At atmospheric pressure this temperature is 573°C. ; it rises slightly with increasing pressure. All quartz is now in the low (α) condition: it may have crystallized originally in this form in quartz veins and in some pegmatites; or it may have inverted from the high form subsequently to its primary crystallization. It has been discovered that there is a measurable and very significant difference between primary α -quartz and inverted β -quartz. On reheating the former, inversion occurs approximately one degree above the temperature at which the latter inverts. The material used in these experiments included crystals from cavities in limestones which are primary α -quartz (known to be so by their crystallographic character), and Hexagonal bipyramidal β -quartz phenocrysts from rhyolites. When subjected to the same heat-treatment quartz from granites behaved in some cases like the former, but in others, like the latter. The significance of this is discussed in the chapter on granites.

Despite the many interesting and indeed unique crystallographic and electrical properties of quartz, the optical characters exhibited in thin sections are of a somewhat negative character. Its refractive indices are low (1.553 and 1.544),¹ with consequent absence of surface relief; it has no cleavage, and the birefringence (0.009) is weak. It even lacks distinctive alteration products, for it is completely stable. These details apply to both the α - and β -phases, and there are, in fact, no definite tests available for differentiating between the two in thin section. Both phases of quartz are liable to contain inclusions of several different kinds, sometimes in such quantity as to render the crystals opaque or at least "milky." Frequently the inclusions are minute gas- or fluid-filled cavities (Fig. 42) often lying in curved planes that evidently represent resealed fractures. Each fluid cavity may contain a gas bubble, mobile in some cases, or even a minute apparently cubic crystal. In some instances the cavity is bounded by plane surfaces giving a *negative crystal* of minute size but perfect form. In other cases acicular needles of rutile are embedded in the quartz. In the granitic rocks termed charnockites, they are exceedingly minute and occur literally in myriads. They produce an optical blue colour in the crystals in which they occur. In smaller numbers they can often be seen, in some Scottish granites, for example; but they are so minute that it is only by analogy with the much stouter prisms that sometimes occur, that they can be identified as rutile.

¹ Because quartz is the only common rock-forming mineral which has a constant and unvarying composition, its refractive indices and birefringence provide valuable standards for comparison.

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Tridymite is the form of SiO_2 stable above 870°C . At its best it forms Orthorhombic tabular plates of almost perfect hexagonal shape; but more commonly occurs as twinned groups of minute crystals, in fine-grained rhyolitic rocks, the best known British examples occurring in the Tardree rhyolite, where the tridymite lines small cavities. Optically tridymite differs from quartz in its

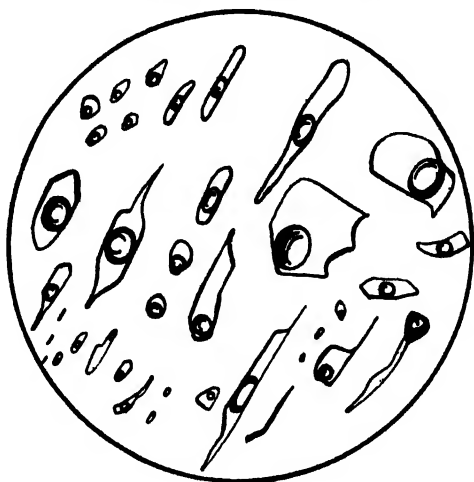


FIG. 42

Fluid cavities containing gas bubbles in α -quartz—highly magnified.

lower refractive indices (1.473 and 1.469) and weaker birefringence (0.004); but its most distinctive characters are the narrow, lath-like crystals and the twinning.

Tridymite has been synthesized using only water as a flux. Two phases are known: tridymite-M (a metastable form) and tridymite-S (the stable form). The former is converted into the latter by prolonged heat-treatment. Both forms have been recognized in silica bricks and igneous rocks of rhyolite-type.

Cristobalite, stable above $1,470^\circ \text{C}$., has long been known from its development in silica bricks used for lining furnaces. Since the advent of X-ray analysis, however, it has proved to be much commoner than was formerly believed in several siliceous igneous rocks. Cristobalite is an important constituent of spherulites in rhyolitic lavas, for example from the Yellowstone Park.¹ The mineral is Cubic, and in certain Hungarian rhyolites cristobalite forms minute crystals, 30 to 60 microns in diameter, in parallel growths, with twinned crystals and radial aggregates lying between the spherulites, which, it is suggested, are themselves high-cristobalite. Somewhat unexpectedly,

¹ Howard, A. D., *Amer. Min.*, 24 (1939), p. 485.

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it has been discovered in the form of half-millimetre cubes and rare cube-octahedrons on the fracture surfaces of olivine-basalt from Plumas Co., California; while it has been recorded from the ground-mass of certain Japanese basaltic lavas. Finally, silica-glass of problematical origin occurring in the Libyan Desert contains small spherulites of cristobalite up to a millimetre in diameter.¹

Both tridymite and cristobalite have lower densities and less closely spaced structures than quartz, so that their formation is favoured by conditions of low pressure.

Cryptocrystalline Silica.—There are many varieties of very finely crystalline silica ranging from radially disposed acicular crystals (fibro-chalcedony) to cryptogranular aggregates difficult to resolve even with high magnification. They are variously coloured, and although the general name chalcedony covers all varieties, such names as carnelian, bloodstone, prase and chrysoprase are used for richly coloured specimens which are used for ornamental purposes. These forms of chalcedony do not occur as primary minerals, but as secondary products they are widespread. The best-known mode of occurrence is as vesicle- and geode-infillings. The relationship of chalcedony – chalcedonite is a better mineral name – to quartz is uncertain. Chalcedonite differs from quartz in the details of its optical properties. The sign of elongation of the fibrous forms varies, so that possibly more than one mineral species is involved. Nevertheless, X-ray studies have revealed no essential structural difference between quartz and “chalcedony.”

Amorphous silica occurs in two natural forms: as **opal**, familiar in its precious form, of course, though common opal, colourless in thin section, and having the same mode of occurrence and origin as chalcedonite, occurs as a secondary product in a variety of rocks. It is identified by its low refractive index, which is somewhat variable according to the amount of water which it contains, but averages 1.445: thus it exhibits a distinct negative surface relief. Normally it is isotropic, but sometimes optical anomalies, particularly very weak birefringence, may be observed. Secondly, **silica-glass** in a number of forms, though a rarity, has been collected from widely-scattered localities. Some forms such as moldavite are apparently of meteoric origin, but other types probably arise as a consequence of the impact of a meteorite, the heat generated being sufficient to cause widespread fusion of superficial deposits, particularly desert sands. Fulgurites also consist of silica-glass forming thin-walled tubes penetrating into sand and formed by fusion by lightning.

¹ Spencer, L. J., “Tektites and silica-glass,” *Min. Mag.*, 25 (1939), p. 436.

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It is impracticable to draw a hard-and-fast line between these two categories of minerals as some play a dual role. Thus the important chlorites are seen to have replaced such mafic silicates as augite, hornblende and biotite in some rocks and are therefore indubitably secondary and replacive; but in other rocks chlorite is evidently a primary constituent, though of late formation.

Indeed in a few instances we have done less than justice to the status of some minerals included in this chapter as accessories, for, although this is their normal role, occasionally they attain to the status of essential minerals. Thus garnets are encountered as accessories in a number of igneous rocks, but eclogite is a two-mineral rock consisting of omphacite and red garnet only, so that the latter is very much an essential. Similarly apatite and sphene are widespread as accessories; but a unique rock-type consisting solely of sphene set in a matrix of apatite occurs in an alkali-complex in the Kola Peninsula, U.S.S.R.

THE SPINEL GROUP

One of the most important groups of accessories comprises the spinels, including magnetite and chromite. The former especially is ubiquitous in igneous rocks of widely varying compositions. Chromite is restricted to coarse-grained ultrabasic rocks and locally occurs in exploitable quantities. In addition to these quantitatively important members of the group there are other spinels which are occasionally encountered in Basic and Ultrabasic rocks. Several of these bear distinctive names such as picotite and pleonaste, but specific identification is impossible without chemical analysis, and generic identification only is expected of the student.

The spinels crystallize in the holosymmetric class of the Cubic System, and it is rare to find forms other than the octahedron represented, though occasionally the edges of the latter are truncated by the rhombododecahedron (110). Twinning on the "Spinel Law" is not uncommon. In this the twin plane is parallel to an octahedron face; the twins are usually flattened parallel to the twin plane, so that the

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crystals have the aspect of flattened triangular tablets, with re-entrant angles at the corners.

The spinels are double oxides of a divalent and a trivalent element respectively, the former being Mg'' , Fe'' , or more rarely Mn'' or Zn'' , and the latter Al''' , Fe''' or Cr''' . The type spinel is the semi-precious gemstone, ruby spinel, $MgO.Al_2O_3$; this is of metamorphic origin and is therefore not considered further. In **pleonaste** some Fe in both states of oxidation occurs, the composition being indicated by $(Mg'', Fe'')O.(Al''', Fe''')_2O_3$.

Picotite is more complicated still, by the addition of Cr''' . This mineral is sometimes therefore named "chrome spinel," which is not synonymous, it should be carefully noted, with chromite. Chromite is $FeO.Cr_2O_3$, while magnetite (described under the heading of "iron ores") is $FeO.Fe_2O_3$.

So far as optical characters are concerned, there is little to say: the shapes seen in thin section are square, triangular or hexagonal according to the attitude of the crystal in the thin section. Refractive indices are very high; and the sections are, of course, isotropic. Magnetite is invariably opaque; chromite is nearly so; but other spinels may be deep green, coffee-brown or plum coloured.

Chromite, as already noted, is found only in olivine-rich ultrabasic igneous rocks such as peridotites or serpentinites derived from them. It crystallizes at high temperatures and is therefore normally euhedral. In an ultrabasic magma effective saturation with chromite occurs early in its cooling history, and a swarm of small octahedral crystals is precipitated, and by reason of their high specific gravity these tend to accumulate in the basal parts of the intrusion, on the then existing floor, where they may form a nearly pure concentrate. Generally, however, packing of the octahedra is not sufficiently close as to exclude other minerals completely, and the latter form the matrix in which the chromite crystals are embedded. Such rocks are termed **chromitites** and are considered in detail below. The identification of chromite in a rock section depends largely on the prior recognition of the rock-type. Large quantities of opaque or nearly opaque octahedral crystals embedded in bronzite together with abundant olivine may safely be identified as chromite: magnetite, the other possibility, would be out of place in this setting.

THE IRON ORE GROUP

The opaque accessories, which include magnetite as well as the other oxide- and sulphide-minerals occurring in igneous rocks, can be adequately studied only with the aid of a metallurgical, as distinct from a petrological, microscope. Identification depends upon colour, reflectivity and optical reactions under *reflected* light. These

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properties are just as distinctive for this class of mineral as are the familiar optical properties of transparent minerals viewed by *transmitted* light. This is not the place to describe the techniques involved; but if the apparatus is available the student is strongly advised to master the techniques, since these are not inherently difficult and are well worth while, if for no other reason than that the opaque minerals display textural relationships which are both interesting and significant in the full understanding of the crystallization history of a given rock. In the absence of the necessary apparatus the common ore minerals may be identified by illuminating the upper surface of the slide, at the same time diverting the transmitted light.

The iron ores that occur as constituents of igneous rocks are magnetite, titanomagnetite, ilmenite, pyrite and rarely pyrrhotite. The first three are normal accessories in a wide range of rock-types, and in some cases occur in such amounts that they must be regarded as important essential components.

Magnetite is the most widely distributed accessory in igneous rocks: it is the "iron ore" of countless petrographic descriptions.

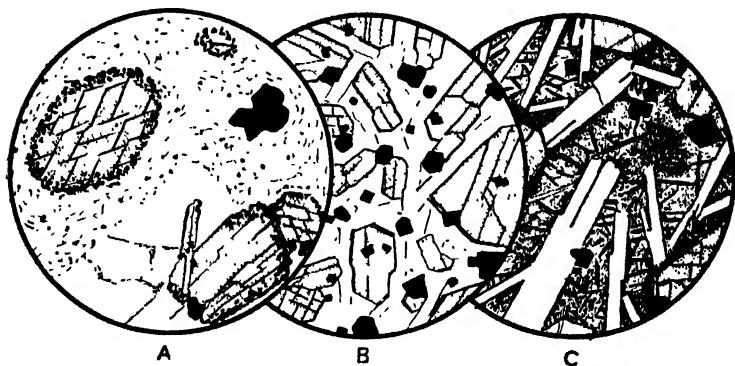


FIG. 43

Iron ores in thin section. (A) Magnetite grains produced round hornblende phenocrysts by magmatic corrosion, in hornblende-andesite, Siebengebirge ($\times 32$); (B) Early crystallized, euhedral magnetite grains in dolerite ($\times 32$); (C) Early formed euhedral crystals, and very late iron ore embedded in glassy groundmass, in olivine-basalt, Hawaii ($\times 60$).

Magnetite is invariably opaque, even in the thinnest sections, and in the absence of crystal form, it may not be easy to distinguish it from other opaque ores; but by oblique reflected light it has a characteristic steely metallic sheen. It is the most strongly magnetic of the iron ores, and may be separated from the others by means of a magnet.

In many igneous rocks, especially the more basic, magnetite is an accessory of early formation—the characteristic, often perfectly

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formed octahedrons occurring plentifully in thin sections of basalts and other lavas, are of this type; but in some basalts it is one of the latest constituents to crystallize, and forms intricate dendritic growths in the interstitial glass (Fig. 43C). Further, magnetite is produced at different stages in the history of a rock by alteration of iron-bearing silicates. A familiar instance is the serpentinization of olivine; while the conversion of amphibole into pyroxene also liberates magnetite (Fig. 43A).

The occurrence of magnetite in layered basic complexes is described later, under "Monomineralic Rocks."

Ilmenite.—Ideally ilmenite is titanite of iron, FeTiO_3 , but in naturally occurring specimens some substitution of Ti by Fe has taken place, so that up to 30 per cent of Fe_2O_3 may be shown on analysis. Ilmenite crystallizes in the Trigonal System, but crystals are rare, and normally the mineral is massive. It is quite opaque, and is therefore sometimes difficult to distinguish from magnetite, especially when quite fresh. Alteration renders the task easier, however, for ilmenite is progressively converted into **leucoxene**. In an early stage, the change is superficial only, and the grains appear white by reflected light. In an advanced stage of alteration the mineral becomes translucent and light brownish-grey, while the Trigonal symmetry is emphasized by thin parallel black bars crossing in three directions (Fig. 44). At one time leucoxene was thought to be secondary sphene in the process of formation, but it has been shown¹ to be hydrated oxide of titanium, $\text{TiO}_2 \cdot n\text{H}_2\text{O}$.

Compared with magnetite, ilmenite is less variable in mode of occurrence. It normally occurs as an accessory mineral, particularly in the more basic, coarse-grained igneous rocks of the gabbro type. In such rocks, indeed, ilmenite is an essential constituent, and occasionally increases in amount, particularly towards the base of layered intrusions. In this country the Carrock Fell gabbro is noted for its richness in ilmenite.²

Having described the properties of magnetite and ilmenite it is necessary to consider the close association of these two minerals, particularly in coarse-grained basic igneous rocks. The nature of the association can be fully appreciated only after examining polished specimens with the metallurgical microscope, when it is commonly seen that apparently single crystal grains are in fact aggregates consisting of magnetite and ilmenite in widely varying proportions. The grain boundaries are generally angular or smoothly curved. In addition to such aggregates of discrete crystal grains of the two ores, either may contain inclusions of the other. They take the form of

¹ Edwards, A. B., "The Chemical Composition of Leucoxene . . .," *Min. Mag.*, 26 (1942), p. 273.

² Harker, A., *Natural History of Igneous Rocks* (1909), p. 133.

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very thin plates or lamellae, the directions of which are determined by the crystal structure of the host mineral. Thus ilmenite lamellae in magnetite are parallel to the octahedron faces of the latter, and on a polished surface form a graticule of either two or three sets of lamellae according to the orientation of the crystal. Magnetite lamellae in ilmenite, however, are confined to the single direction,

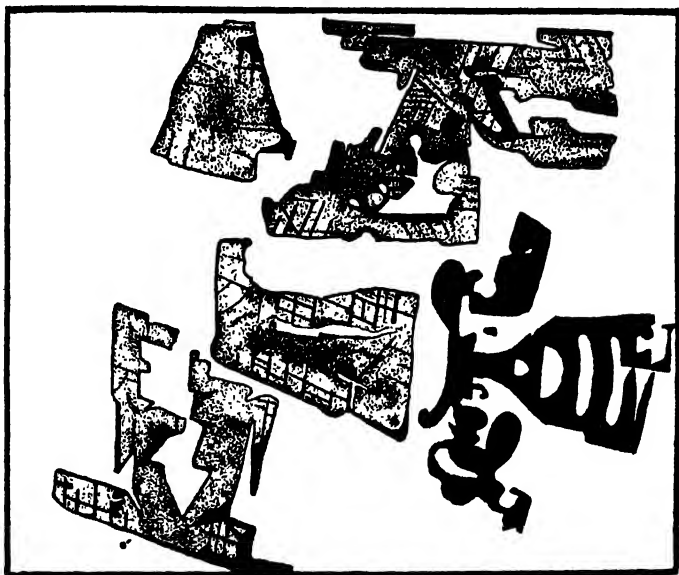


FIG. 44

Ilmenite and its alteration products; from lugarite, Lugar, Ayrshire. The one skeletal crystal is fresh; the others have been altered into grey-brown translucent leucoxene traversed by bars of iron-oxide.

parallel to the basal pinacoid (0001) of the ilmenite. These intergrowths are exactly analogous with the perthitic intergrowths in feldspars and pyroxenes, already described, and like them are the result of unmixing during slow cooling.¹ This conclusion is suggested by the textural relationships between the two ore minerals; it is confirmed by the fact that in many basalts, which are the high-temperature equivalents of gabbros, only a single, homogeneous iron-ore phase occurs—and is a titaniferous magnetite or **titanomagnetite**.

Ilmenomagnetite is the unmixed equivalent of titanomagnetite as described above. Other oxide minerals including haematite and spinel enter into the ilmenomagnetite but only in very small quantities. They are visible only under high magnification, and their

¹ Buddington, A. F., *et al.*, "Thermometric and Petrogenetic Significance of Titaniferous Magnetite," *Amer. Journ. Sci.*, **253** (1955), p. 497.

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identification is really a matter for the specialist. If haematite is involved in the intergrowths it is generally as a secondary mineral replacing original magnetite, particularly around grain boundaries and along fractures.

Haematite, Fe_2O_3 , is a Trigonal mineral well represented in collections by the fine, brilliantly lustrous crystals obtained from iron-ore deposits. Haematite in this form does not occur in igneous rocks; but it is quite common as minute specks disseminated through, or coating other minerals and sometimes penetrating into them in the form of thin veinlets. Haematite is opaque in the mass and gives a whitish steely reflection from polished surfaces; but when it occurs as an incrustation or a thin film it may be translucent and of a brilliant red colour. Even minute amounts of haematite act as pigment, and the pale pink to brick-red colours so common among feldspars are largely a result of this pigmentation.

Limonite is a comprehensive name given to a complex group of hydrated oxides of iron with the general formula: $\text{Fe}_2\text{O}_3 \cdot n\text{H}_2\text{O}$. These minerals are always secondary, formed at low temperatures in the presence of water, as weathering products resulting from the decomposition of iron-bearing minerals. Limonite itself is amorphous; but within the group of hydrated ferric oxides one mineral in particular achieves a minutely crystalline condition. This is **goethite**, $\text{Fe}_2\text{O}_3 \cdot \text{H}_2\text{O}$, which, although showing a general resemblance to limonite, can be distinguished from it by its anisotropic optical properties, including pleochroism and birefringence. The precise identification of the generally minute amounts of yellowish-brown secondary iron ore in igneous rocks is difficult and of little practical significance: it is normally regarded as adequate to record such material as "limonite."

Pyrite is the most widespread sulphide mineral, FeS_2 , which crystallizes in one of the lower symmetry classes of the Cubic System. The forms exhibited include the cube, octahedron, and occasionally the very distinctive pentagonal dodecahedron or pyritohedron (210). The crystallographic characters, light brassy colour and metallic lustre render pyrite distinctive.

In thin section it is opaque, but the distinction from magnetite and ilmenite should present no difficulty, for the metallic lustre and brassy appearance are diagnostic when the mineral is examined by reflected light.

It is doubtful if pyrite ever occurs as a primary constituent of igneous rocks, but it is common as a secondary mineral, associated with calcite, chlorite and secondary quartz, etc.

Pyrrhotite has the same chemical composition as pyrite, but crystallizes in the Hexagonal System. When quite fresh pyrrhotite resembles pyrite, though it is somewhat darker in colour. Exposure

ACCESSORY AND SECONDARY MINERALS

causes it to tarnish, and ultimately it assumes a characteristic bronzy appearance. A further point of difference between these two sulphides is that pyrrhotite is magnetic, though variably so. Pyrrhotite is an important accessory in certain basic igneous rocks, it has been recorded from some nepheline-bearing rocks and, rarely, from granitic rocks. The best-known occurrence is the Sudbury "norite" in Ontario, in which the pyrrhotite appears to be a mineral of late crystallization, although it is believed that the sulphidic material separated at an early stage from the silicate-melt as a consequence of immiscibility. This interpretation is based largely on the mode of occurrence of the pyrrhotite, much of which has the form of composite globular aggregates which are interpreted as immiscible droplets.

A British example at the other end of the compositional range is the Shap Granite, in which pyrrhotite is an accessory: the opaque grains seen in thin sections of this well-known rock are pyrrhotite—not magnetite.

In favourable circumstances pyrrhotite may be used in geological thermometry: it is the high-temperature form of FeS_2 , and it is therefore important to differentiate between it and pyrite.

THE ZIRCONIUM-BEARING ACCESSORIES

Zircon, a silicate of zirconium, ZrSiO_4 , is one of the best-known minerals which crystallize in the Tetragonal System. The crystals

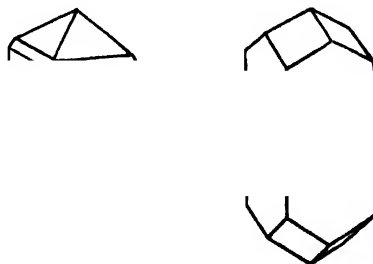


FIG. 45

Crystals of zircon.

Left, vertical prism (110) combined with Tetragonal bipyramids (111) and (331);
Right, prism (100) with bipyramid (111).

consist of bipyramids in combination with prisms, sometimes the form (110), sometimes the form (100) being dominant. Again, the bipyramid faces may be in zone with the prism, or may be of a different order (Fig. 45).

The crystals are commonly light brown, buff or reddish-brown,

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while phenocrystic zircons of bright red colour, with subvitreous lustre, occur in a grey slaggy lava from Niedermendig, Germany.

Zircons are common in granitic, syenitic and dioritic rocks, but appear to be rarer in those of more basic composition. As a general rule they are of small (microscopic) size, colourless with very high refractive indices ($\alpha = 1.927$, $\gamma = 1.982$). The birefringence is likewise strong (0.055). As they crystallize at a high temperature, they are liable to be enclosed in minerals of later formation.

In certain types of nepheline-syenites the zircons are of much larger size and have the status of an essential constituent, particularly in the coarse-grained "zircon-syenite-pegmatites." Zircons are interesting from several points of view: they are important accessories in many igneous rocks, and frequently in the systematic separation of the heavy minerals during the study of an igneous rock a "flood" of perfectly formed small zircons is apparent. These crystals are in the main colourless and transparent; but purple, yellowish or brownish translucent crystals may also occur.¹ By reason of its radioactivity zircon was the mineral used in pioneer attempts to estimate the age of igneous rocks, using for the purpose biotite-granites which contain zircons surrounded by the well-known *pleochroic haloes* when embedded in dark mica. The possibility that careful study of shape-variation of zircons might throw light on petrogenetic problems, particularly when the possibility of replacement is involved, has been realized by some petrologists; but the results are difficult to interpret. In some cases rocks of igneous aspect have been found to include rounded zircons, presumably of sedimentary origin, which might be taken to imply that the rock-body has been produced by *in situ* replacement of the country rock; but it might equally well mean that some zircon-containing sediment had been incorporated in magma and completely assimilated except for the zircons which are chemically stable. On the other hand the several members of a great batholythic complex have been shown to contain euhedral zircons which evidently crystallized at high temperatures as normal accessories in an igneous environment. The crystal habit was found to be constant within each member of the complex, but they were distinctive individually, reflecting slight differences in the cooling conditions.²

Eudialyte is one of a number of complex zircono-silicates³ which occur as accessory minerals in nepheline-syenites, particularly of

¹ The highly prized blue zircons of gemstone quality do not occur naturally in this colour: they are heat-treated.

² Larsen, L. H., and Poldevaart, A., "Measurement and distribution of zircons in some granitic rocks . . .," *Min. Mag.*, 31 (1957), p. 544.

³ The composition of eudialyte is indicated by the following formula, based on a recent analysis:

$(\text{NaCa})_5\text{ZrSi}_6\text{O}_{17}(\text{O},\text{OH},\text{Cl})$, with 0.6 per cent of $(\text{Nb},\text{Ta})_2\text{O}_5$.

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pegmatitic facies. It increases in amount in more basic nepheline-bearing rocks and may attain to the status of an important essential constituent, as in the so-called eudialyte-syenites in southern Greenland. One rock near Julianehaab, in this area, contains eudialyte to the extent of a third of the whole rock. It is also important in similar rocks in the Kola Peninsula, U.S.S.R. In Britain it has been recorded from quartz-syenite veins in limestone at Barnavave, Ireland.¹

Eudialyte is a Trigonal mineral. It is commonly bright red with vitreous lustre in the hand specimen, and therefore very distinctive in appearance. In thin section it is often colourless, but may be pink, and then exhibits a pink to yellow pleochroism. Refrindex is moderate, with $\alpha = 1.609$ and $\gamma = 1.611$. The birefringence is notably weak—about 0.002.

TITANIUM-BEARING ACCESSORIES

(i) **Rutile, Anatase and Brookite.**—Rutile is the most stable of the three crystalline forms of TiO_2 —rutile, anatase and brookite—and is the one which most commonly occurs as an accessory mineral, especially in the more acid rocks. It forms slender Tetragonal prisms and needles which are frequently twinned to give geniculate and heart-shaped forms. Both in hand-specimens and under the microscope, rutile is coloured yellow or red. It is remarkable for its exceptionally high refractive indices (2.61 and 2.90), the highest for any of the normal rock-forming minerals.

Rutile occurs in crystalline schists and is also common in sands and sedimentary rocks. Fine hair-like crystals occurring in clay and slates (clay-slate needles) are referred to rutile. In pneumatolytized granites, rutile is often present as a product of the breakdown of complex titanium-bearing minerals, such as biotite and sphene. Under these circumstances it may form an interlocking structure of regularly orientated needles, known as sagenite. Quartz crystals in certain granites contain myriads of exceedingly minute needles believed to be rutile, by analogy with the megascopic, reddish-brown acicular crystals of more robust habit occurring in the quartz of certain pegmatites. Relatively large crystals also occur in eclogites.

Anatase, the second of the TiO_2 minerals, differs from rutile and brookite in crystallographic characters and optical properties. It crystallizes in the holosymmetric class of the Tetragonal System, and exhibits two different habits: (1) tabular, due to the strong development of the basal pinacoid, often modified on the edges or corners by feebly developed bipyramid faces; and (2) steep bipyramids, usually with well-developed surface markings (Fig. 46).

In thin section anatase is scarcely ever seen in ordinary rocks; but

¹ Nockolds, S. R., *Min. Mag.*, 29 (1950), p. 27.

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in granitic sands containing a natural concentrate of heavy constituents it is often plentiful. The anatase tablets vary from colourless to rather dull shades of blue, green or brown. The refractive indices are very high (2.49 and 2.56), and the birefringence is also strong (0.07). In view of its distinctive appearance a confirmatory test is seldom necessary, but the tablets show a uniaxial (positive) interference figure in convergent light.

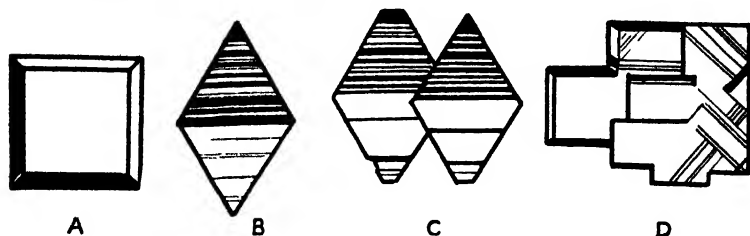


FIG. 46

Crystals of anatase: (A) and (D) of tabular habit, with characteristic surface markings in the latter case; (B) and (C) of bipyramidal habit, and, in the latter case, two crystals in parallel growth.

Brookite crystallizes in the Orthorhombic System, in platy crystals with the pinacoid (100) well developed, and showing strong vertical striations by reason of oscillation with the unit prism.

Again we turn to granitic sands, like those from Dartmeet on Dartmoor, to study typical crystals or grains. They are light yellow to brown in tint, with closely ruled striations as noted above. The really distinctive feature, however, is the failure to extinguish, on account of crossed axial-plane dispersion. The effect of this property is to cause the section to display a succession of interference colours, in the order of Newton's scale, when the stage is rotated with both polarizer and analyser in operation.

(ii) **Sphene** or **Titanite**.—Sphene is a silicate of titanium and calcium, CaTiSiO_5 , occasionally seen in mineral collections as small wedge-shaped crystals, but much better known as a distinctive and widespread accessory mineral in rocks of many different kinds. Sphene can sometimes be seen even with the naked eye, and easily with a lens, in hand-specimens of granodiorites, syenites and diorites, for example. The crystals, which belong to the Monoclinic System, are brilliantly lustrous and range in colour from light yellow to green or brown.

In thin section sphene may be colourless, but is commonly a shade of greyish-brown, and is slightly pleochroic. The refractive indices, 2.01 and 1.90, are notably high, so that surface relief is strong. The birefringence also is particularly strong, so that the interference colours are very light—the so-called “high order whites”—and

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these, superimposed on the normal absorption tint of the mineral, make no appreciable difference. Thus, provided the section is not in the position of extinction, crossing and uncrossing the polarizers makes no difference to the colour of the section. This observation provides a unique test for the identification of sphene.

As an accessory in igneous rocks sphene normally occurs as small, scattered crystals; but a rock-type has been discovered in the Kola Peninsula, U.S.S.R., consisting of crowded crystals of sphene embedded in a matrix of apatite—a truly extraordinary rock-type.

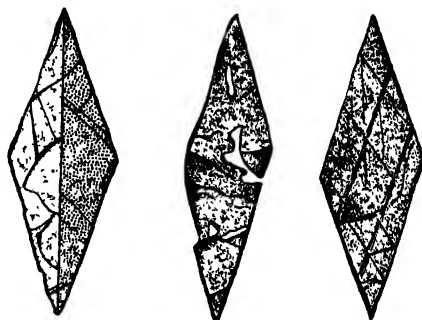


FIG. 47

Sections of sphene: two on right from ditroite, Ditro, Transylvania, one showing cleavages, the other corrosion inlets; on the left a simple twin from leucite phonolite, Perlerkopf, Brohlthal.

In different rocks sphene may show very different relationships towards the minerals with which it is associated. As a normal accessory it is common and easily recognized (Fig. 47). Much less commonly it is moulded upon the minerals in which it is normally embedded: instead of being euhedral it is xenomorphic, and in a dioritic assemblage is bounded by the plane faces of plagioclase and amphibole.¹ This sphene is primary, but belongs to a late stage in the crystallization sequence. Finally, sphene may occur in association with other secondary minerals pseudomorphing one of the primary constituents. For example, certain dioritic rocks from Jersey, Channel Islands, contain pseudomorphs after euhedral Ti-rich amphiboles consisting of penninitic chlorite containing much sphene, the latter representing the titanium and calcium which could not be incorporated in the chlorite.² In the rare rock, borolanite, garnets occur which contain appreciable amounts of sphene

¹ and ² Wells, A. K., and Bishop, A. C., "An appinitic facies . . . in Jersey, Channel Islands," *Q.J.G.S.*, 111 (1955), 143-66, especially Figs. 5, 10 and 12.

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of almost the same colour as the melanite in which it is embedded. Evidently in this case after the initial crystallization of a Ti-rich garnet, presumably schorlomite, owing to a change in environmental conditions it became unstable and inverted to a Ca- and Ti-poor garnet intimately associated with sphene.

(iii) **Perovskite**.—Among the rarer accessory minerals perovskite fills an important role in some uncommon rock-types. It is calcium titanate, CaTiO_3 , and crystallizes in the Cubic System, the octahedron being the only form normally developed. The small crystals are isotropic, and Cubic beyond question; but larger crystals exhibit optical anomalies and, it is recorded, quite considerable birefringence. The outstanding optical property is a very high refractive index (2.38), which makes the small crystals stand out strongly, despite their size. In fact very small crystals have such a heavy margin that they appear almost opaque. Cleavage is stated to be perfect, parallel to cube faces.

Perovskite is characteristic of, and restricted to, under-saturated lime-rich rocks, and is a constant associate of melilite with which it occurs as small grey octahedrons and twinned groups. Probably the most notable occurrences so far described are in ultrabasic lavas from Uganda. The perovskite in these lavas occurs in unusually large quantities, up to 6.2 per cent in one type. In these rocks the perovskite is golden-brown, yellow or green instead of the commoner grey.

APATITE

Apatites are essentially phosphates of calcium, with small amounts of fluorine, chlorine, and/or hydroxyl. According to the dominance of either fluorine or chlorine, two varieties, fluor-apatite and chlor-apatite are distinguished: the former is much the commoner in rocks. Apatite crystallizes in the Hexagonal System, in the class in which there are no vertical planes of symmetry; but usually the crystal development is so simple—a combination of prism in zone with a bipyramid and basal pinacoid—that this passes unnoticed. Apatite may be colourless, but is often bluish, brown or green in hand specimens. Commonly it occurs as minute euhedral crystals of prismatic habit which may be seen in thin sections of most igneous rocks. Relatively large crystals occur in coarse-grained syenites, especially syenite-pegmatites; at the other extreme are hair-like “needles” scarcely visible under ordinary magnification. Although normally colourless in thin section they may appear bluish, and quite commonly have cloudy coloured cores.

Apatites may be recognized by their form, moderate relief (refractive indices about 1.63), very low-order weak-grey interference

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colours and isotropic basal sections. Crystals are often fractured across by a basal parting.

Beryl is a silicate of beryllium and aluminium which crystallizes in the holohedral class of the Hexagonal System, and is the mineral generally used to demonstrate the crystallography of that class, since beautiful crystals, obtained from granite pegmatites, are available. Beryl is familiar, too, in gemstone form, particularly the bluish-green aquamarine, the brilliant emerald and the golden variety. Some beryls of phenomenal size have been mined from certain pegmatites; very much smaller crystals are found occasionally in cavities in granites, including those of the Mourne Mts., in north-eastern Ireland and Lundy Island. It is worth noting that even the smaller crystals represent a remarkable feat in terms of natural concentration of beryllium. The latter is one of many so-called "trace-elements," which may exist in rocks in concentrations of only a few parts per million. Normally these minute amounts are accommodated in the structures of common minerals; but in beryl and other, even rarer, pegmatite minerals, the trace-elements are sufficiently concentrated to form independent compounds.

GARNET GROUP

Garnets are silicates of aluminium, iron, manganese, chromium, calcium and magnesium. The crystal structure consists of separated SiO_4 groups, and there are three of such groups in the unit of the structure, giving Si_3O_{12} which appears in all the formulae. In this structure there are twelve free links to which other atoms are attached according to the scheme:

$\text{M}''_3\text{R}'''_2(\text{Si}_3\text{O}_{12})$, where M is a divalent, and R a trivalent element. The following varieties may be distinguished:

	<i>Formula</i>	<i>Colour</i>	<i>Refr. Index</i>
Pyrope	$\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$	blood red	1.705
Almandine	$\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}$	dark red	1.830
Spessartine	$\text{Mn}_3\text{Al}_2\text{Si}_3\text{O}_{12}$	red	1.800
Uvarovite	$\text{Ca}_3\text{Cr}_2\text{Si}_3\text{O}_{12}$	green	1.870
Grossularite	$\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$	green or orange-red	1.735
Andradite			
(Melanite)	$\text{Ca}_3\text{Fe}_2\text{Si}_3\text{O}_{12}$	black	1.895

These are largely hypothetical: natural garnets are more complex in composition than any of these "pure" garnets and in any specific

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case may be regarded as containing *either* the first three, *or* the second three components in varying proportions. Thus a common red garnet on analysis would be found to contain Mg'' , Fe'' and Mn'' , and according to the amounts present might be expressed in terms of the molecular percentages of Py, Al and Sp respectively. Both groups of three form continuously variable series, but grossularite and pyrope, for example, mix only with difficulty under high pressure.

Crystallographically the garnets appear to be simple: the only forms occurring normally are the rhombdodecahedron (110), the icositetrahedron (211) and much less commonly the hexakis octahedron (321). Garnets of the Py.Al.Sp. type are isotropic in thin section, and there seems no reason to doubt their full Cubic symmetry; but those in the second group frequently exhibit optical anomalies. The examination of the optical characters suggests that the apparently simple crystals are actually complex twins consisting of twelve or twenty-four individuals, which are biaxial and may be Orthorhombic. Above $800^{\circ}C.$ they assume Cubic symmetry.

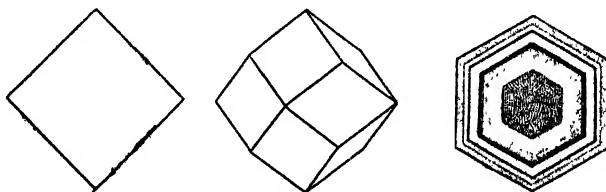


FIG. 48

Garnet: *Centre*, the simplest form, the rhombdodecahedron (110); *left*, a section through centre of crystal, parallel to one of the axial planes of symmetry (100), (010), or (001); *right*, section through centre of crystal parallel to one of the twelve dodecahedral faces. Zoning is formally shown.

In thin section all garnets are characterized by high relief, on account of the high refractive indices; but the figures quoted above must be regarded as approximations only, on account of the complex composition of natural garnets. As might be expected, the sections often exhibit zoning, reflecting changes of composition during growth (Fig. 48). No cleavages occur in garnets in rocks; but detrital garnet separated from sands frequently shows a good dodecahedral cleavage developed as a consequence of severe pounding by wave action. Garnets alter in a variety of ways. Thus in rocks which have suffered contact metamorphism, garnet may be pseudomorphed by cordierite. The commonest alteration product is probably chlorite; but garnet carrying titanium (schorlomite) may alter in part into sphene.

Undoubtedly the finest crystals of garnet occur in, or have been obtained from garnet-mica-schist. But in addition to this—the

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commonest—mode of occurrence, garnet may occur as a normal accessory in a wide range of igneous rocks. In granitic dyke-rocks a garnet of the first group, often rich in manganese, occurs in this way. Syenitic rocks containing feldspathoids such as nepheline and leucite, not infrequently contain garnets near to melanite (andradite) in composition. As might be anticipated, accessory garnets are found in ultrabasic igneous rocks such as peridotites, and are rich in magnesium. They tend to survive, as distinctive red crystals, when such rocks are converted into serpentinites (see below). Grossularite has an entirely different mode of occurrence, and is almost restricted to calcareous sedimentary rocks which have suffered thermal metamorphism. Andradite, too, is characteristic of such an environment, and is prominent in the so-called skarns—lime-silicate rocks from intrusive contact-zones.

THE PNEUMATOLYTIC MINERALS

Tourmaline.—Tourmalines are complex hydrated silicates and borates of aluminium, magnesium and sodium, with iron, manganese, calcium and small quantities of potassium, lithium and fluorine. The tourmalines may, according to their constituent bases, be classified as alkali-tourmalines, iron-tourmalines, magnesium-tourmalines, etc. They all contain about 10 per cent of boric acid, and from $3\frac{1}{2}$ to 4 per cent of water. Tourmaline crystals are Trigonal, occurring in triangular prisms with pyramidal terminations, sometimes accompanied by a basal plane or pedion.

From many points of view, tourmaline is a mineral of outstanding interest. Crystallographically it provides striking examples of polar symmetry, the terminations of some specimens being obviously different; while if the crystals are so simple as to show a single horizontal plane at both ends, the complete independence of these pedions is proved by physical differences between them, one being as lustrous as glass, while the other may be lustreless or strongly etched. This polar symmetry is in many instances shown by an unsymmetrical distribution of colour along the length of the crystal: this is most spectacular in certain parti-coloured crystals of gemstone quality, which are coloured rich pink at one end and bright green at the other.

Characteristically the prism zone is strongly striated due to oscillation between the dominant trigonal prism and the hexagonal prism which modifies its edges (Fig. 49).

In thin section tourmaline may occasionally be colourless; but normally it is blue, brown or green, and strongly pleochroic. The absorption of the ordinary ray is much stronger than the extraordinary, so that prismatic sections show their deepest colour when

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the principal axis lies at right angles to the vibration direction of the polarizer in the microscope.¹

Basal sections, often of modified triangular cross section, are frequently zoned, even if the crystal is minutely acicular. They yield a negative uniaxial interference figure in convergent light. The

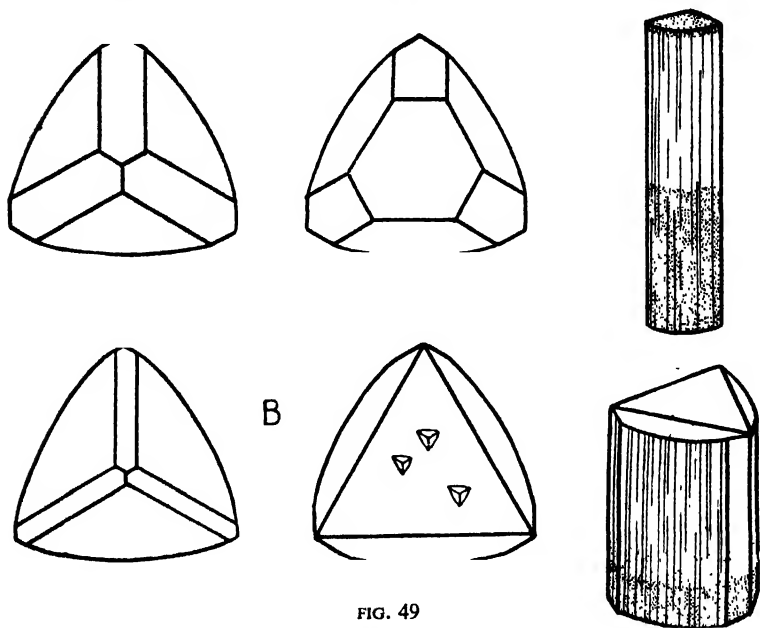


FIG. 49

Tourmaline crystals: (A) The two differing terminations of one crystal, showing combinations of positive and negative trigonal pyramids at one end, and similar forms with the addition of a pedion (0001) at the other. (B) Similar combinations shown by a crystal of slightly different habit, with the characteristic shape of etched figures added diagrammatically. The crystals (right) show polar colour variation and vertical striations. Tourmalines of gem quality from Minas Geraes, Brazil.

birefringence of tourmaline varies with composition, reaching a maximum of 0.046 in varieties rich in chromium, but averaging 0.029 in "schorl"—the type commonly occurring in tourmaline-granites and associated rocks.

Tourmaline is essentially a component of the so-called pneumatolytic rocks and of granite-pegmatites. Thus it is widely distributed in and around the West of England granites. In part it is a primary mineral, but in large measure it has been produced at the expense of pre-existing mica and feldspar, in which case it commonly

¹ The vibration of the polarizer lies parallel to the cleavage traces in a vertical section of biotite, when the stage has been turned to the position in which the mineral shows its deepest absorption tint.

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assumes a distinctive habit, much-elongated acicular crystals with a radial disposition, or being so closely packed as to form a felt-like aggregate (see under "Pneumatolysis," and Fig. 79).

Topaz is fluosilicate of aluminium, $\text{Al}_2(\text{F},\text{OH})_2\text{SiO}_4$, and crystallizes in the holosymmetric class of the Orthorhombic System. Typical crystals are illustrated in Fig. 50.

Topaz of gemstone quality from certain well-known localities in Brazil has a distinctive honey-brown colour; but other varieties

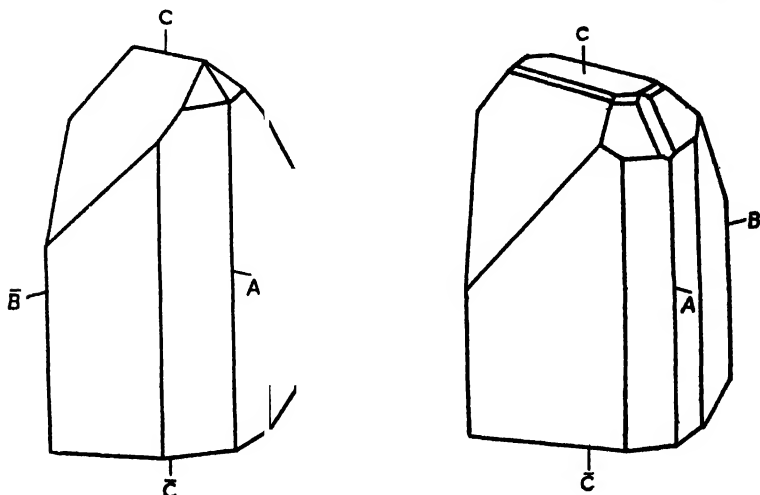


FIG. 50

Crystal habits of topaz from Kleine Spitzkop, South-West Africa.

Combinations of vertical prisms (110) and (210); the dominant "dome" (021), the "dome" (101) and bipyramids. The crystals are single ended, being bounded below by basal cleavage planes.

may be faintly tinted blue, though many are quite colourless. The single cleavage, parallel to the base of the crystal, is perfect, though difficult to produce artificially.

Optically topaz is characterized by moderate refractive indices, 1.615 for α and 1.625 for γ . Thus the birefringence, 0.009, is almost identical with that of quartz, though the relief is considerably stronger. Naturally the mineral is biaxial; it is of positive sign, with the acute bisectrix emerging perpendicular to the basal cleavage. Thus an observation in convergent light will usually confirm the diagnosis.

Topaz occurs in irregular grains and spongy masses in pneumatolytic rocks, notably in greisens. In such rocks the amount of topaz may be very large: in a "topazfels" from Schneckenstein in Saxony,

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80 per cent of the rock is topaz. The mineral also occurs rarely as a primary mineral in granites, though in Britain it appears to be restricted to irregular druses in the Mourne Mountains and Lundy Island granites. The large crystals of gemstone quality are obtained from pegmatites.

Fluorite (popularly termed fluorspar)

Fluorite is the fluoride of calcium which crystallizes in the holosymmetric class of the Cubic System. Normally the cube, as simple crystals or interpenetrant twins, is the only form seen; but occasionally the cubes may be modified on the edges by the tetrahedron (210) or on the corners by the octahedron (111). Fluorite may be quite colourless, but is commonly tinted green, blue, or typically violet. A most distinctive feature is the brilliant fluorescence in vivid, luminous blue, when illuminated with ultraviolet light.

In thin section fluorite shows very strong negative surface relief (on account of its abnormally low refractive index, 1.434), and traces of a highly perfect octahedral cleavage (Fig. 80). The sections are perfectly isotropic. Normally they are colourless but occasionally the characteristic violet colour appears, even in a section of normal thickness, and often the colour is patchily distributed.

Fluorite occurs in pneumatolytic rocks and is especially characteristic of greisens, where it is associated with topaz. It is also a common gangue mineral, for example in the lead-mining area of the northern Pennines, whence come so many of the beautifully crystallized specimens that grace mineral-cabinets.

CONTAMINATION ACCESSORIES

Corundum

There are three very distinct varieties of this mineral—common corundum, and the clear gem-stones, ruby and sapphire. These all crystallize in the rhombohedral class of the Trigonal System, but the crystal habit is very variable: the crystals are combinations of hexagonal prism, bipyramid and basal pinacoid, with the rhombohedron, the presence of which establishes the symmetry class to which the mineral belongs, and in addition determines the directions of the characteristic surface markings on some of the crystal faces (Fig. 51C). Oscillation between the prism and the bipyramid causes strongly developed horizontal striations (Fig. 51B). It is probably more than a coincidence that true sapphires are frequently of bipyramidal habit (the clear blue crystal shown in Fig. 51, A is a case in point); while rubies of gemstone quality are often rhombohedral. Common corundums may be tabular, bipyramidal with very irregular faces (Fig. 51B) or prismatic (Fig. 51C).

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In composition all varieties are essentially alike, only differing in the minute amounts of trace elements that act as pigments to the coloured varieties. Apart from these, corundum is just crystallized alumina, Al_2O_3 .

Corundum is notable for its extreme hardness (9 in Moh's scale). It has a parting parallel to the (0001) face. Optically corundum is distinguished by its high refractive indices ($\alpha = 1.760$, $\gamma = 1.768$) and weak birefringence (0.008), almost the same as quartz. Normally corundum is colourless in thin section, but a strongly coloured

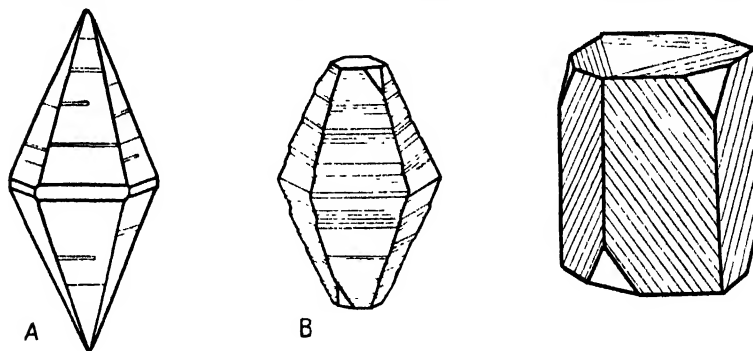


FIG. 51

Three crystal habits of corundum.

- (A) Sapphire: combination of hexagonal prism and bipyramid.
- (B) Opaque red corundum, Ceylon: combination of heavily striated bipyramid, rhombohedron and basal pinacoid.
- (C) Common corundum, Pietersburg, Transvaal, South Africa: combination of hexagonal prism, rhombohedron and basal pinacoid, with characteristic surface markings.

sapphire is light blue, and of course slightly pleochroic. In convergent light a uniaxial negative interference figure confirms the identification.

Although corundum is typically a product of the thermal metamorphism of argillaceous sediments, it does originate in other ways, and is a rare constituent of some types of igneous rocks. In the form of small blue sapphires, corundum has been described from argillaceous xenoliths in basic intrusions of Tertiary age in Mull and Ardnamurchan. Naturally, corundum can occur only in quartz-free igneous rocks, and it appears to be restricted to those rocks which have suffered desilication: that is, they have contributed silica to the adjacent wall-rock. Thus certain syenitic and dioritic rocks from the Bancroft area in Ontario and from some South African localities, are corundum-bearing.

Andalusite.—This mineral is one of a number of silicates of aluminium and may be represented by the formula Al_2SiO_5 , being

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thus identical in composition with sillimanite (which may be regarded as the high-temperature equivalent of andalusite), and kyanite, its stress equivalent. Andalusite is an Orthorhombic mineral, but it exhibits strong pseudo-tetragonal symmetry, and unless the crystals are terminated, they might easily pass for Tetragonal prisms (Fig. 52). In appearance the crystals are rather undistinguished: they are usually grey, with resinous lustre, and a surface coating of white mica into which the mineral readily alters.

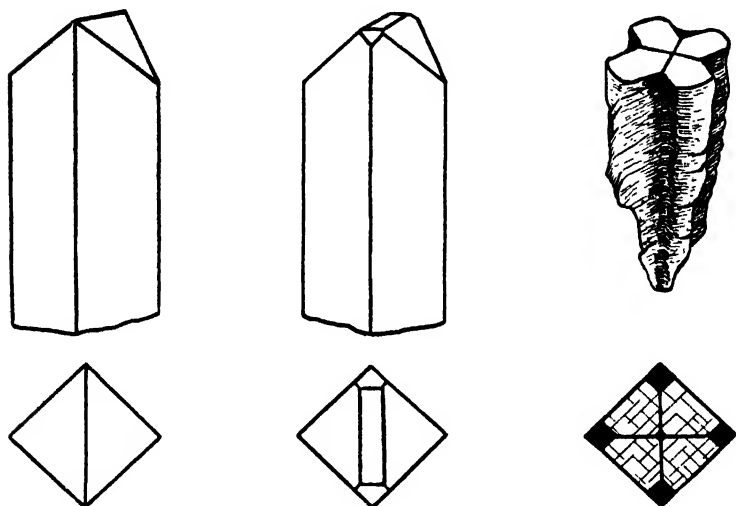


FIG. 52

Andalusite and chiastolite crystals.

Left: Combination of prism (110) and "dome" (011). *Centre:* A more complicated crystal showing the forms (001) and (101) in addition. *Right:* A weathered crystal of chiastolite. *Below:* The ideal cross-section showing prismatic cleavages and characteristic carbonaceous inclusions.

In thin section andalusite may be distinctive, largely by reason of its pleochroism. At its best the colour corresponding to the slow vibration, Z, is peach pink, while X and Y may be both colourless, or there may be a trace of bluish-green. This pleochroism scheme is, in fact, closely similar to that of hypersthene. In many rocks, however, the andalusite is non-pleochroic. The refractive indices are moderate, with $\alpha = 1.635$ and $\gamma = 1.643$ (average values); while the birefringence is weak, varying in different specimens from a little below to a little above quartz. Traces of prismatic cleavage, in basal sections crossing at almost a right angle, are consistently developed, but the sections are strongly sieved with small inclusions

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of the associated minerals, which sometimes render systematic examination rather difficult.

Typically andalusite occurs in argillaceous hornfelses, in which it is accompanied by cordierite, biotite and often spinel, and is one of the first newly crystalline minerals to appear in these rocks.

It also occurs somewhat rarely as an accessory in igneous rocks, and must then be regarded as evidence of the assimilation of argillaceous xenoliths. Thus the Armorican granites of the West of England in places contain andalusite especially in contaminated facies.

Chiaustolite is merely a variety of andalusite containing regularly arranged carbonaceous inclusions (Fig. 52). It occurs in the so-called chiaustolite-slate within the aureole of thermal metamorphism round granites. Well-known British examples occur near the contact between the Skiddaw granite and the Skiddaw slates.

Cordierite

This silicate of aluminium, magnesium and iron occurs in a wide variety of rocks of different kinds. It is often a challenge to the skill of the petrographer on account of lack of distinctiveness in its optical characters. The composition may be represented by the formula $Mg_2Al_4Si_5O_{18}$; but this may be written in a form reminiscent of beryl, except that the Si_5O_{18} of the latter becomes $(AlSi_5)O_{18}$ —a proxy-Al replacing a Si atom—in cordierite. This is significant, for although cordierite is an Orthorhombic mineral, it is strongly pseudohexagonal, particularly when twinned. Natural crystals are prismatic in habit, but are relatively rare. When fresh, cordierite is violet in colour and pleochroic, the obsolete name “dichroite” having been applied to it on account of its pleochroism, which is, however, not seen in sections of normal thickness. There is no true cleavage, but a basal parting is characteristic.

The optical properties vary with the content of iron; but commonly the sections are quite colourless in thin slice, while the indices are low— $\alpha = 1.535$, $\gamma = 1.544$. Thus one index is the same as quartz and the surface relief is the same as that of the latter mineral, so is the birefringence (0.009); and it is therefore often very difficult to distinguish between the two minerals. Under the most favourable conditions the twinning is diagnostic. Ideally the twins comprise three or six sectors; but actually the number of sectors varies (Fig. 53), and it may be complicated by lamellar twinning superimposed on the sectorial. As the twin planes are prism faces, vertical sections are similar to simple twins of feldspar, while the lamellar twinning may cause an uncomfortably close resemblance to plagioclase. However, cordierite is characteristically closely associated with spinel, often in swarms of small dark green octahedra; and yellow pleochoric haloes occur round minute inclusions of zircon and rarer

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accessories. Finally, the crystals may show any degree of alteration to an aggregate of white mica and chlorite, termed "pinite." Obviously as the mineral is Orthorhombic it must be biaxial—a fact readily checked in a basal section, to which the acute bisectrix, X , is perpendicular.

As regards mode of occurrence, cordierite is typically a product of thermal metamorphism of argillaceous sediments. It occurs, also, though much less commonly, in basic and in acid igneous rocks, in which it undoubtedly represents imperfectly digested xenoliths of argillaceous material. The cordierite-garnet-norites of Aberdeenshire provide good examples.¹ The Land's End granite contains

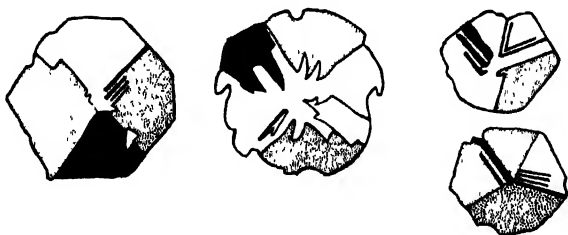


FIG. 53

Sections of cordierite between crossed polarizers,
showing sectorial twinning.
From cordierite-norite, Arnage, Banffshire.

ehedral greenish-grey pinitic pseudomorphs after cordierite, up to an inch or so in length. More rarely the mineral is found—again usually in xenoliths—in lavas such as basalt and andesite. Cordierite crystals or aggregates, often of large size, occur also in pegmatites in several localities. As the mineral is not usually present in the wall-rock, which in different instances is igneous (either acid or basic) or metamorphic, it is inferred that the cordierite has been formed from the pegmatitic liquid, *not* from caught-up (xenolithic) material.

THE ZEOLITES

The zeolites constitute a group of hydrated silicates of aluminium with Na' , K' , Ca'' and more rarely Ba'' , so that they are closely comparable in composition with the feldspars and feldspathoids. They are commonly formed from the latter by hydrothermal alteration; and conversely, on heating under conditions of thermal metamorphism zeolites alter very readily into feldspars of the appropriate composition. There are more named zeolites than there are

¹ Read, H. H., "The Geology of Banff, Huntly and Turriff," *Mem. Geol. Surv.* (1923), pp. 128-37.

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species in any other mineral group so far dealt with; but in spite of their numerical superiority, with one exception they are of minor significance among the rock-forming minerals.

Their chief interest is centred in their crystallographic features; but unfortunately space does not allow these to be enlarged upon. Many zeolites are pseudosymmetrical: they appear, on superficial examination, to belong to one crystal system; but careful measurement shows them to belong to another, of lower symmetry. Thus crystals of apophyllite often appear to be a simple combination of the cube and octahedron, with a pearly lustre on some of the "cube" faces; but actually the crystals are stumpy Tetragonal crystals—pearly faces which catch the eye are basal pinacoids, the triangular "octahedral" faces are bipyramids. Several species form fibrous aggregates, each "fibre" being a much-elongated prism. In hand specimens these look much alike; but they may be identified optically. Other zeolites are easily identified by their crystallographic features, as detailed in mineralogy textbooks.

Zeolites may be grouped in several ways for descriptive purposes—according to their chemical composition or crystallographic characters for example; but we prefer to deal with them in natural associations of zeolites as encountered in basaltic lavas.

The most important mode of occurrence of the zeolites is as infillings of vesicles, geodes and irregular cavities in lavas, notably basalts (p. 136). Many fine specimens are collected from such occurrences, for example at the Giant's Causeway in Antrim. Systematic studies of the distribution of zeolites in the Tertiary volcanic region of Antrim¹ has brought to light a zonal arrangement of different associations, on the largest scale within the lava field as a whole, and on a small scale within a single lava flow; while there are significant differences in the zeolite-assemblages in lavas of different petrographic types. These assemblages define "zones" which are mappable and which were evidently dependent upon the temperature involved. Zeolitization is believed to have been caused by the circulation of water, largely of meteoric origin, and heated during a period of vulcanicity. Lower zeolite zones reached a higher temperature than higher zones, but pressure also was involved, for it is believed that the zeolites were deposited under a cover of 500 to 1,000 feet of overlying lavas.

To emphasize the zonal relationship the zeolites may be described as they occur in these natural associations.

1. The **chabazite-thomsonite** association. The former' zeolite occurs as aggregates of Trigonal crystals of rhombohedral habit, looking almost like cubes, while thomsonite is one of the fibrous

¹ Walker, G. P. L., "The amygdale minerals in the Tertiary lavas of Ireland," *Min. Mag.*, 32 (1960), p. 503.

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zeolites and one of the most widely distributed. It is a "soda-lime" zeolite and one which might be expected to result from hydrothermal alteration of plagioclase.

2. The **stilbite-heulandite** association. Although these two minerals may occur separately, they are usually seen in close association in beautifully crystalline aggregates. Their most striking physical character is a pearly lustre resulting from a single cleavage direction, parallel to the side pinacoid. Stilbite characteristically forms composite crystals, pinched in at the middle giving a general resemblance to a wheatsheaf. Heulandite, on the other hand, usually forms single perfect Monoclinic crystals of a coffin shape. The stilbite-heulandite association occurs typically in the reddened top, highly vesicular part of olivine-basalt flows, the central more compact parts containing thomsonite instead.

3. The **natrolite-analcite** association. These two minerals of similar composition define a zone, though they are more or less mutually exclusive. Natrolite is a fibrous zeolite, essentially sodic and therefore closely similar to albite and nepheline. When natrolite occurs in sufficiently robust crystals, these look to be simple Tetragonal prisms capped by the unit bipyramid, but the ratio between the "a" and "b" axes is 0.978 : 1 so that the crystals are actually Orthorhombic.

Analcite is the most important zeolite to the petrologist as it is the only one which attains to a rock-forming status. Chemically it may be represented by the formula $\text{NaAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$ though some substitution by K' for Na' is liable to occur. Phenocrysts of potassium-analcite have been described from lavas in Montana, while leucite has been pseudomorphed by analcite elsewhere.

Analcite crystallizes in the Cubic System and is superficially holosymmetric, the only common form being the icositetrahedron, though rarely the cube may occur in addition. These features may be deceptive, however, and analcite may be another case of pseudosymmetry: in some thin sections anomalous double refraction occurs and complex twinning effects similar to those seen in leucite may be observed. The reader will realize that analcite and leucite have many features in common; shape, general appearance, colour and lustre are closely similar; but the modes of occurrence are completely different (though see porphyritic analcite above). Leucite forms scattered phenocrysts in lavas. Analcite, in well-formed crystals, commonly lines cavities in highly altered rocks: it forms aggregates rather than single crystals, which are usually off-white and opaque, but from some localities are tinted pink.

In thin section analcite may be difficult to identify with certainty, on account of its rather negative character. Typically it is colourless and transparent, of low relief (R.I., 1.487). Careful mani-

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pulation of the iris diaphragm may reveal traces of cubic cleavage, and make it easy to apply the Becke test, which will confirm its low refractive index. The optical anomalies visible between crossed polarizers have already been mentioned. In altered rocks analcite may be replaced by a finely crystalline aggregate of usually unidentifiable material.

In British rocks analcite is widely distributed especially in the Intermediate and Basic intrusives of Carboniferous age, notably *teschenites* and *crinanites* and in basaltic lavas of comparable composition. The mineral is probably most abundant in the so-called analcite-syenite, an example occurring at Howford Bridge, Ayrshire.

The associates of analcite in these rocks are ilmenite, labradorite, lilac-coloured titanite, and red-brown barkevikite. In all these rocks the analcite is a primary mineral, though of late formation. Therefore it normally occurs in interstitial patches between the earlier formed constituents; but with increasing amounts it spreads into the adjacent plagioclases along cleavages and veins, and progressively replaces them. This *analcitization* is a late-stage replacement comparable in its effects with albitization.

THE EPIDOTE GROUP

This group of minerals comprises the Orthorhombic *zoisite*, and the Monoclinic *clinozoisite*, *epidote* (*pistacite*), *piedmontite* and *orthite* (*allanite*).

The composition of *zoisite* may be represented by the formula $(\text{OH})\text{Ca}_2\text{Al}_3\text{Si}_3\text{O}_{12}$. If any iron replaces aluminium, the substance tends to crystallize in the Monoclinic System as *clinozoisite*. The latter is one end member of a continuous series, the other being the hypothetical "iron-epidote" $(\text{OH})\text{Ca}_2\text{Fe}_3\text{Si}_3\text{O}_{12}$. If the latter does not amount to more than 10 per cent of the whole, the mineral is *clinozoisite*; if between 10 and 40 per cent, the term *epidote* is applied. No natural *epidote* contains more than 40 per cent of the iron-bearing component. Two members of the group are notably rich in manganese: *thulite* is a manganiferous *zoisite*; while *piedmontite* is a manganiferous *epidote* with the formula $(\text{OH})\text{Ca}_2(\text{Al}, \text{Mn})_3\text{Si}_3\text{O}_{12}$. The variety *withamite* is a poorly manganiferous *piedmontite*. Finally *allanite*, sometimes called *orthite*, is essentially a variety of *epidote* containing the rare elements cerium and yttrium.

Clinozoisite-Epidote.—All Monoclinic members of the group occur in crystals elongated along the *b*-axis, and difficulty is experienced in sorting out the faces in the principal zone parallel to this direction. Commonly the basal pinacoid (001), orthopinacoid (100) and a hemiorthodome (101) occur in this zone, and while some of the faces are strongly striated in characteristic fashion, others may

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be perfectly smooth and highly lustrous, while again others may be relatively dull. Two observations may be of assistance: the common type of twin, often betrayed by a re-entrant angle at the ends of the crystal, has the twin-plane parallel to (100), which may therefore be identified; while the cleavage is parallel to the basal pinacoid (001).

In its simplest development the end of the crystal of epidote shows the unit prism, with faces meeting in an edge parallel to the twin plane and to the face (100). Sometimes the crystals are terminated by a hemibipyramid; or by combinations of prisms and hemibipyramids.

Crystals of epidote may be nearly black, but usually are of a distinctive shade of yellowish green, termed pistachio green, hence the synonym, pistacite. The darker coloured crystals may bear a strong superficial resemblance to tourmaline; but, unless they are too thick, may be readily distinguished by the different absorption of light in different directions perpendicular to the principal crystallographic *b*-axis of epidote. Tourmaline, of course, can show no such differences in this zone.

Optical Properties

The optical and other physical properties vary systematically from clinozoisite to ferriferous epidote. The minimum and maximum refractive indices and the birefringence are tabulated below:

	clinozoisite	ferriferous epidote
α	1.725	1.732
γ	1.740	1.781
birefringence ($\gamma - \alpha$)	0.015	0.049

Thus the highest interference colours in a slide of normal thickness (30 microns) should be a pale third order yellow for iron-rich epidote. The value of the birefringence rises rapidly with increasing iron content. When the latter is low, the sections are colourless; but increasing iron causes a light yellow colour and slight pleochroism. A noteworthy feature of epidote and other minerals in the group is the noticeable variability in birefringence even within the limits of a small crystal. The birefringence of clinozoisite is notably anomalous, on account of strong dispersion: the interference colours are rich dark-blue or brown, like those of penninitic chlorite.

The plane of optical symmetry is perpendicular to the length of the crystal, therefore all sections lying with the principal crystallographic *b*-axis in, or parallel to, the plane of the slide, will exhibit straight extinction. Sections cut perpendicular to the length, however, show oblique extinction up to 30 degrees measured between the slow vibration direction (*Z*) and the traces of the (001) cleavage.

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The optical orientation of epidote is such that one optic axis emerges at right angles (approximately) to the (100) face, on which a detrital grain or small crystal will normally lie; while the other optic axis is almost perpendicular to the basal cleavage. Therefore the distinctive optic axis figure appropriate to such a section is seen in convergent light both in crystals of normal habit and in cleavage flakes.

Allanite or **Orthite** is also distinctively coloured in thin section, and strongly pleochroic, from buff to deep red-brown.¹ Twinning, usually simple, is common on the usual epidote plan. Identification of allanite is aided by its common association with another member of the epidote group—sometimes even in zonal association.

Mode of Occurrence and Origin

Excluding zoisite, all the chief members of the group are produced by, and characteristic of, dynamothermal metamorphism, in the course of which both coloured silicates and the anorthite-component of plagioclase break down into new stable associations, prominent among which is a member of the epidote group. Plagioclase may be represented by perfect pseudomorphs consisting of packed aggregates of zoisite prisms in slightly divergent groups; or the central parts of a large crystal may be composed of similar crystals or granular aggregates, embedded in, and surrounded by, albite.

The clinozoisite molecule may be regarded as a high-density representative of anorthite. Obviously, unless there has been an actual influx of Fe ions, anorthite can be replaced only by zoisite or clinozoisite (typically the latter). Similarly, both common augite and hornblende contain the components of epidote, and the latter is produced, usually in association with chlorite, under the conditions that result in the replacement of anorthite. This type of replacement is often termed **epidotization**.

In addition to this secondary development, *primary* epidote occurs in some pegmatites, and less frequently even in granites and other normal igneous rocks.

In many other rocks occurring in areas which have *not* been subjected to regional metamorphism, epidote or its varieties may be common, having been formed by late-stage (deuteric) readjustments. The substance of this epidote has been derived from the mafic minerals, including hornblende and augite, both of which are often pseudomorphed by penninitic chlorite. The unwanted Ca⁺⁺ ions (together with the other necessary components) go to form epidote embedded in the chlorite.

¹ Owing to the strong absorption, it is often difficult or impossible to check the character of the vibration, whether fast or slow, in the usual manner using a quartz wedge.

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THE CARBONATES

The carbonates fall into two groups for descriptive purposes, the one comprising those which crystallize in the Rhombohedral Class of the Trigonal System, and the other, much less important as rock-formers, which are Orthorhombic. The latter do not normally occur in igneous rocks and are therefore omitted from this account.

The rhombohedral carbonates with their chemical formulae and certain of their optical properties are tabulated below:

		γ	α	$\gamma - \alpha$
Calcite	CaCO_3	1.658	1.486	0.172
Magnesite	MgCO_3	1.700	1.509	0.191
Siderite	FeCO_3	1.875	1.633	0.242
Dolomite	$\text{CaMg}(\text{CO}_3)_2$	1.680	1.501	0.178
Ankerite	$\text{Ca}(\text{Mg,Fe})(\text{CO}_3)_2$	—	—	—

Analyses show that there is a considerable degree of miscibility between the various end-members, and also with MnCO_3 (rhodochrosite).

Calcite shows a greater variety of crystal habits than any other mineral. The combinations of forms include hexagonal prisms, basal pinacoid, acute, unit and obtuse rhombohedra, and the very characteristic "general form", the scalenohedron (21 $\bar{3}$ 1).

The characteristic and almost sole form exhibited by the other members of the group is the unit rhombohedron; but commonly the faces of the crystals are composite and strongly curved, giving them a very characteristic saddle-shaped appearance.

Apart from their crystallographic characters, the outstanding feature of the rhombohedral carbonates is their perfect cleavage, parallel to the faces of the unit rhombohedron (10 $\bar{1}$ 1).

In thin sections it is impracticable to differentiate between the several members of the group by ordinary methods, but it is relatively easy to do so using microchemical tests which are described in the appropriate textbooks on petrographic methods. As is shown in the above table, the double refraction is outstandingly strong. Therefore in sections of normal thickness they polarize in high order colours and often appear dappled with very "watery" pink and green. In calcite, one of the indices is below, the other well above that of Canada balsam, so that any section of the mineral (other than the basal one which is, of course, singly refracting), changes in appearance as it is rotated over the polarizer. In certain positions the refractive index of the light passing through it will equal that of the balsam, so that the calcite practically disappears. When turned through a right angle, however, the outline becomes bold

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and the surface relief strong. Thus the rapid appearance and disappearance of strong surface relief on rotation of the stage constitutes a unique test for calcite.

The carbonates are characteristic weathering products of igneous rocks, particularly the more basic ones. In extreme cases little but the original texture remains: all the component minerals have been replaced by calcite, with small amounts of other secondary minerals. Normally calcite arises by weathering or hydrothermal alteration of calcium-rich silicates, notably the more basic plagioclases. In addition, calcite is a very common associate of zeolites, chlorite, chalcodony, etc., in vesicles and amygdules.

In two groups of rocks the presence of carbonates is of special significance. In some nepheline-syenites calcite occurs and has all the appearance of a primary mineral: it occurs in anhedral grains surrounded by and embedded in other primary minerals, that thus appear to be of later formation. This fact is significant in connection with the problem of the origin of this type of igneous rock, and is more fully considered in due course. Intrusive veins, dykes and more particularly plugs of crystalline carbonate rocks occurring in close association with nephelinic rocks forming ring-complexes are of special interest, and their origin is a major petrological problem. The intrusive carbonate rocks are termed carbonatites: they are described and discussed in a later chapter.

The only other igneous rocks which contain carbonates in significant amounts are certain lamprophyres in which calcite may be sufficiently abundant to cause effervescence on treatment with dilute acid. This calcite is of late-stage, deuteric origin, and frequently pseudomorphs the primary silicates.

THE CHLORITE-SERPENTINE GROUP.

The minerals falling in this group are often regarded as belonging to two groups, the chlorites and the serpentines; but all these minerals are closely related in composition and in physical characters, so it seems best to consider them together. Treated thus, the serpentine minerals become end-members of the larger group. Fundamentally, they are hydrous silicates of aluminium and magnesium, with iron tending to replace both of the latter components. Actually the composition of an individual specimen is complex, but can be expressed in terms of the end-members of a variable series, the more important being:

Antigorite (a serpentine) $(\text{OH})_4\text{Mg}_3\text{Si}_2\text{O}_5$, which is comparable with Ferroantigorite, in which all the Mg ions are replaced by Fe.

Amesite, in which Al replaces part of the Si, with the necessary adjustment of the valency— $(\text{OH})_4\text{Mg}_2\text{Al}_2\text{SiO}_5$; while

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Daphnite is the corresponding iron-bearing member:
 $(\text{OH})_4\text{Fe}_2\text{Al}_2\text{SiO}_5$.

Winchell names thirteen members of the group, each with a definite range of composition, but their identification is a problem for the expert mineralogist. Therefore the present account is limited to the essential characters of the commoner members of the group.

As the name implies, the chlorites are typically deep bluish-green in colour, and often superficially resemble the micas, though they are softer (hardness, 2 to 3); they have a soapy "feel" in hand-specimens; and while they cleave readily, the cleavage flakes are not elastic like those of mica. Crystallographically also they resemble the micas in being actually Monoclinic but with pseudo-hexagonal symmetry.

Again like the micas, the chlorites are of the sheeted layer type as regards atomic structure, which accounts, of course, for their perfect basal cleavage. The fundamental basis of the structure is a sheet of SiO_4 -tetrahedra linked into a repeating hexagonal pattern, like that illustrated for the micas in Fig. 22. The arrangement of the atomic sheets is polar, however, the stack illustrated in Fig. 23 being repeated below the horizontal plane, instead of being reflected across it, as in the micas.

X-ray studies of the serpentine minerals have shown that serpentine rocks ("serpentinites") consist of three minerals essentially the same in chemical composition, $(\text{OH})_4\text{Mg}_3\text{Si}_2\text{O}_5$, but differing in the details of their structure. A rough and ready distinction may be made on the basis of differences in habit: **chrysotile** is fibrous; **antigorite** is lamellar; while **lizardite** is massive. Much of the pale coloured light green or yellowish green mineral in many serpentinites is lizardite.

Bastite is the name given to the serpentine mineral which pseudomorphs orthopyroxene. Generally its identification is quite easy, as enough of the original structure survives to suggest orthopyroxene by comparison with the very different appearance of the pseudomorphs after olivine.

Penninite is one of the commoner chlorites, being a widely distributed alteration product of micas, amphiboles and pyroxenes. Penninite is an iron-poor chlorite, relatively deficient in aluminium, and may be expressed as dominantly $\text{Ant}_{60-80}\text{Am}_{40-20}$. It is the chlorite nearest in composition to the serpentine antigorite. It may be noted that some of the Al ions occupy positions in the structure that are normally filled by silicon: *i.e.*, they lie in the centre of tetrahedral groups of four oxygens, and should be regarded as proxy-aluminums. In thin section the mineral appears light green and slightly pleochroic. Frequently it is fibrous, and shows a radial to

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spherulitic structure. With indices $\alpha = 1.575$ and $\gamma = 1.576$, the birefringence is very weak—0.001. Coupled with strong dispersion, this gives an anomalous interference colour described as “ultra blue.” It is a deep inky blue usually, but deep brown, or rich violet are also sometimes exhibited. As the acute bisectrix, Z , is perpendicular to 001, and therefore to the cleavage traces, all sections showing the cleavage traces will prove to be fast along the cleavages.

Clinochlore or Clinochlorite.—This chlorite has the same limited iron-content as penninite, but is richer in Al. The composition may be expressed as $\text{Ant}_{60-40}\text{Am}_{40-60}$, that is, about equal amounts of antigorite and amesite molecules. The general optical orientation is the same as penninite but clinochlorite may show slightly oblique extinction (2 to 9°) measured to the cleavage traces, hence the name of the mineral. The surface relief is much the same as for penninite, but the birefringence is slightly greater (0.004 to 0.011). The most distinctive feature, however, is repeated twinning on the mica law.

Iddingsite and Bowlingite.—In certain types of basaltic rocks the olivine phenocrysts are partially or completely converted into a strongly coloured yellowish or reddish-brown lamellar substance, known under one of the above names. Their exact composition seems uncertain, but they are easily altered into limonitic pseudomorphs.

Iddingsite has been proved by X-ray analysis to be a complex alteration product, not a single mineral. The only crystalline phase in iddingsite is goethite, a hydrated oxide of iron, the rest is amorphous. Bowlingite may well be similar: it is more strongly coloured than iddingsite, in rich red shades.

Chlorophaeite.—Under this name a strongly coloured chloritic mineral has been described from certain of the basaltic and doleritic rocks of Carboniferous age from the Midland Valley of Scotland. Its most distinctive feature is a rich green colour when fresh; but it is prone to very rapid oxidation when it turns brown.

TALC

This mineral resembles the chlorites and micas in having an atomic structure of the extended sheet type, which imparts the usual perfect basal cleavage and pseudohexagonal symmetry. Since the linkage between the atomic sheets is exceptionally weak, the latter readily slide over one another, and consequently talc is used as a “dry” lubricant. It is exceedingly soft, being the indicator mineral for “hardness 1” on Mohs’ scale; while a soapy feel is a characteristic feature—the popular name for massive talc (steatite) is soapstone.

Talc is a hydrated silicate of magnesium and may be represented by the formula $(\text{OH})_2\text{Mg}_3\text{Si}_4\text{O}_{10}$. It is produced by the alteration of magnesium-rich silicates, and is typically found in metamorphosed

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ultrabasic and basic igneous rocks. One well-known mode of occurrence is the so-called actinolite-talc schist, in which the talc acts as matrix to the actinolite crystals. The mineral occurs, too, in serpentinites, and in these and all other occurrences there is a distinct resemblance to white mica in thin sections. The optical orientation of talc is the same as white mica, with the optic axial plane and the acute bisectrix, X , perpendicular to (001). Thus the sign is negative; but positive elongation is shown in all vertical sections. Further, the refraction ($\gamma = 1.575$ to 1.590 , $\alpha = 1.538$ to 1.545) and double refraction (0.030 to 0.050) are similar to white mica. In spite of these resemblances to white mica, particularly to sericite, there is no risk of confusion, for the associates of the two minerals are entirely different: white mica is characteristic of acid rocks, while, as noted above, talc occurs in basic and ultrabasic assemblages.

PART II

**MODES OF OCCURRENCE OF
IGNEOUS ROCKS
AND
CONSOLIDATION OF MAGMA**

MODES OF OCCURRENCE OF EXTRUSIVE
IGNEOUS ROCKS

IGNEOUS rock-masses may be very simply divided into two major categories, viz.: (i) those that crystallize within the crust; and (ii) those that solidify at the surface. The former are commonly called "intrusions," the latter extrusions, or lava flows.

Since lava flows are the only igneous rock-bodies which can be actually observed in the process of formation, they form the logical starting-point for the study of igneous rocks. They demonstrably result from the consolidation of extremely hot magma erupted at the surface. Only in so far as the intrusive rocks are analogous in composition and texture to the lavas, can a common magmatic origin be legitimately inferred. No question arises as to the origin of most of the sheet-like intrusions (dykes and sills): they evidently represent material identical with that of the lava flows, but injected into crustal fissures. The field relations of certain "igneous" veins suggest a replacive, rather than a displacive origin, however. There remain the coarsely crystalline silicate rocks, often of great surface area, and unknown—and unknowable—extension in depth. These are commonly classified as "major intrusions"; but in the present state of knowledge it is preferable to use a non-committal term with no implications as to origin, such as "pluton."

Lava Flows

Lava is mobile rock material—magma—poured out through a vent or fissure. The flow may be subaerial or submarine. In subaerial flows the lava forms streams of greater or less extent, according to the fluidity of the magma, the more basic being the more fluid. Thus, some of the Icelandic lavas of basaltic composition were so fluid that they extended for forty to fifty miles from the point of eruption.

With submarine flows the lava is spread over, and succeeded by, normal sediment: it forms part of the stratigraphical succession and is said to be *contemporaneous*. The date of the outpouring can obviously be fixed by means of the fossils contained in the contiguous strata. Thus lavas erupted at intervals during the Carboniferous

PETROLOGY OF THE IGNEOUS ROCKS

Period cover considerable areas in the Midland Valley of Scotland. Basaltic outpourings on the grandest scale occupy vast areas in Idaho, Oregon and Washington, in the Parana Basin in Brazil and in the Deccan, India, where the lavas known as the Deccan Traps cover an area of some 200,000 square miles. These vast basaltic plateaux appear to have been built up by numbers of flows, individually of no great thickness, erupted through fissures, and therefore fed by dykes. This form of extrusion, known as **fissure eruption**, has taken place within historical times in Iceland. On account of the relative absence of explosive volcanic activity, fissure eruptions are believed to take place with a quiet outwelling of the lava.

Contrasting with these fissure eruptions are those of **central type**. In these, the energy concerned in producing the uprise of the magma, instead of being dispersed over wide areas, is localized at certain points, where the familiar craters are established. The actual form of the volcano is determined by the type of magma, which is more variable in composition than in the plateau-lavas of fissure eruptions. Where the magma is basaltic, widespread lava-domes of the type magnificently exemplified by the Hawaiian volcanoes, are built up. This volcanic pile, covering an area of well over a hundred miles in horizontal extent at the base, rises from the floor of the Pacific to a height of 13,000 feet above sea level. Contrasted with these lava-domes are the very much smaller, steep-sided cones, built largely of exploded volcanic debris, loosely known as volcanic ash, lying at the angle of rest, and to some extent reinforced by veins of intrusive rock. The well-known volcanoes of Etna and Vesuvius are of this type.

The trachytic lavas of Auvergne in France were erupted in such a viscous condition as to form hummocky or dome-shaped masses of small lateral extent. These are the familiar "puys" of Auvergne. A special development of the latter type is illustrated by the disastrous eruption of Mont Pelée in Martinique, when, following the formation of an incandescent "cloud" (*nuée ardente*) of finely divided ash, a massive "spine" or column of andesitic lava was slowly protruded, and reached a height of 800 feet above the crater.

Volcanic Necks.—A volcano of central type is connected to the underground source of magma by a pipe of roughly circular cross-section. With the cessation of activity this "neck" survives long after all trace of the actual crater and of the lavas erupted from it have disappeared as a consequence of denudation. The material which fills the neck is of two kinds: firstly, it may be lava which congealed before reaching the crater; or it may be fragmental material formed by explosions during the period of activity. The former is covered by the term "*vent intrusion*," and the latter is said to be *pyroclastic*, and may range from accumulations of great angular blocks to finely pulverized rock—the so-called volcanic

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ash. These pyroclastic materials are considered more fully in a later section. The vent-intrusions are usually lithologically and texturally indistinguishable from surface lava flows.

In this country parts of the Midland Valley of Scotland are

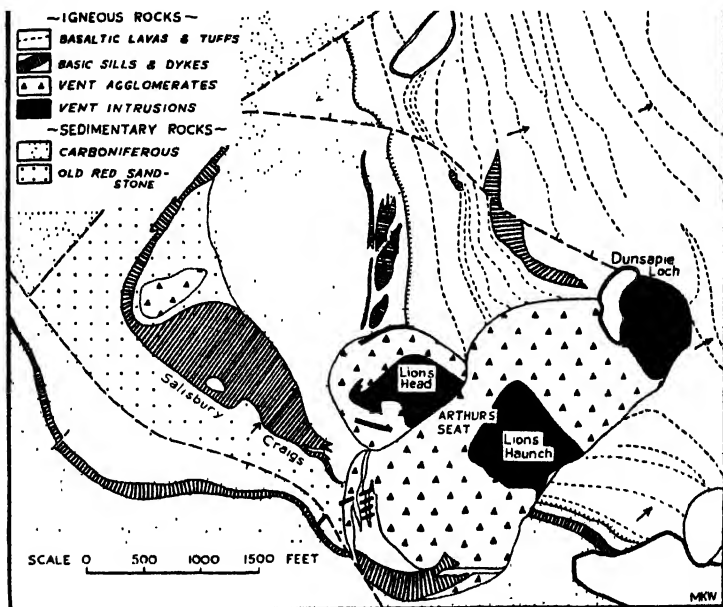


FIG. 54

Simplified map of the Carboniferous lava flows, intrusions and volcanic vents in the vicinity of Edinburgh.

(After H.M. Geol. Survey)

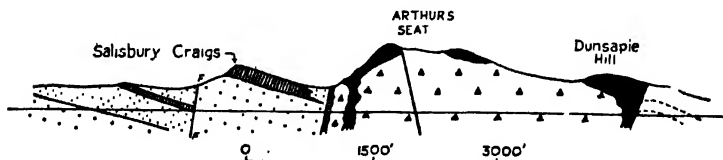


FIG. 55

Cross-section of the Arthur's Seat vents. The key to the ornament is given in Fig. 54.

noteworthy by reason of the large numbers of small volcanic necks which have been located, especially in Fifeshire and Ayrshire. In the former locality many of the vents are seen in cross-section on the coast. An outstanding example, occurring on the outskirts of Edinburgh, forms Arthur's Seat, a very prominent landmark over a wide stretch of country (Fig. 54).

The Structure of Lavas

From direct observation it is possible to distinguish a number of types of lava-flow. Under subaerial conditions the flow may resemble a tumbled mass of clinker or slag. This is distinguished as block-lava, and results when the volatiles in the magma are boiled off in the vent, before eruption. With more rapid uprise and less loss of volatiles the solidified lava shows on the surface contorted, snaky folds, suggestive of irregular viscous flow. This is ropy lava. A feature highly characteristic of submarine lavas of all ages is pillow



FIG. 56

Pillow lavas (spilites) with jasper between some of the pillows, Anglesey.

(M. K. W. after photo by H.M. Geol. Survey.)

structure. Lavas exhibiting this structure consist of isolated pillow-shaped masses piled one upon another, the intervening spaces being filled with sedimentary material, sometimes chert, sometimes limestone, sometimes hardened shale (Fig. 56). Internally the pillows are characterized by concentrically arranged vesicles, and occasionally there is a central ovoid cavity. Pillow structure is shown by basaltic and andesitic lavas, while the most perfect examples occur in flows of spilite. In the case of spilites, instead of parting at once with the steam incorporated in the lava, when erupted on the sea-floor, the pillows swelled up like a sponge and retained sufficient

steam in the vesicles to drastically reduce the density. Moreover, such lavas on eruption seem to have been in the spheroidal state, and since each spheroid or pillow as it became detached was jacketed in steam, the lava flowing over the sea-floor formed a mobile sheet of rolling spheres, seldom touching one another until they had cooled down. Subaerial lavas cannot show typical pillow structure. In Britain the best examples of pillow-lavas are of Precambrian and Ordovician age, and are described in the final section of this book. Basaltic pillow lavas of Tertiary age occur in Mull, western Scotland, at points where basalt flowed into a crater-lake.¹

As a direct consequence of the conditions of outpouring, certain lavas are distinguished by marked flow- (or fluxion-) structure, due to the rolling over and pulling out of the semi-solid mass. This may result in very distinctive internal structures described in a later chapter. Certain lavas are characterized by a cellular or scoriaceous structure due to the elimination of water vapour and other gases during eruption.

Basaltic magma normally has a relatively low viscosity so that upon eruption, when pressure on the magma is reduced, the dissolved gases are largely able to escape into the air. They are in effect boiled out of the lava. However, the loss of volatile substances (which act as fluxes in their dissolved state) causes a rapid increase in viscosity. Many of the gas bubbles thus become trapped by the congealing lava, particularly towards the top of a flow. The gas cavities or vesicles are spherical or ellipsoidal, but may be flattened and elongated by flow movements of the lava. Diameters commonly range from a few millimetres to one or more centimetres, though larger, less regular cavities also occur.

In all recently erupted lavas the vesicles are gas-filled cavities; but in older basalts they are commonly filled-in by low-temperature minerals such as calcite, chalcedony and zeolites. Aggregates formed by these light-coloured minerals are frequently of the approximate shapes and sizes of almonds so that they are called amygdaloids.²

Amygdaloidal basalts are frequently conspicuous among the early members of a lava succession, and it is probable that the amygdale minerals are precipitated from circulating ground-water which becomes heated volcanically under an insulating blanket formed by overlying flows. Statistical studies of the distribution of the vesicle-infilling minerals have demonstrated a zonal distribution within a particular volcanic area, presumably reflecting temperature control. The zones are characterized by different mineral assemblages; while, as might be expected, different types

¹ "Tertiary and Post-Tertiary Rocks of Mull," *Mem. Geol. Surv.* (1924), p. 109.

² Latin: *amygdalus*—almond.

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of lava give rise to distinctive assemblages of vesicle minerals.¹ In Britain fine mineral specimens, particularly of zeolitic assemblages, are obtainable from the lavas of Devonian age in Scotland (Old Dumbarton), and of Tertiary age at the Giant's Causeway, County Antrim.



FIG. 57

Amygdaloidal basalt, Butaure, Tyrol. The amygdales are less regularly shaped than usual and were infilled with various minerals deposited in zonal sequence. In the large, central amygdale calcite forms the outer zone, followed inwards by chlorite and chalcedony showing agate structure.

Pipe amygdales have been described from lavas in many parts of the world. They commonly occur at the base of the flow, and have the form of long narrow cones, tapering upwards, up to a foot in length, and perhaps half an inch in diameter. They are filled with the same kinds of minerals as occur in the more normal ovoid amygdales. They appear to have resulted from the uprise of steam from the moist surface over which the lava flowed, for they commonly occur where lava was erupted over wet mud.²

As a consequence of contraction due to cooling many igneous rocks come to be traversed by regularly arranged systems of intersecting joints. In its most perfect development—in certain basic lavas and minor intrusions—the resulting structure is aptly termed **columnar structure**. The columns, which are very often long and regular, are bounded by three, four, five or six planes, producing triangular, quadrangular, pentagonal, and hexagonal prisms. Where

MODES OF OCCURRENCE OF EXTRUSIVE IGNEOUS ROCKS

the rock-texture is homogeneous the six-sided prisms are most prevalent, for of all the cases in which the centres of contraction are equidistant, and the angles of the prisms fit together without any intervening space, the hexagonal arrangement gives the highest ratio of area to periphery. The long axes of the columns are perpendicular to the retreating isotherms during cooling. In accordance with this

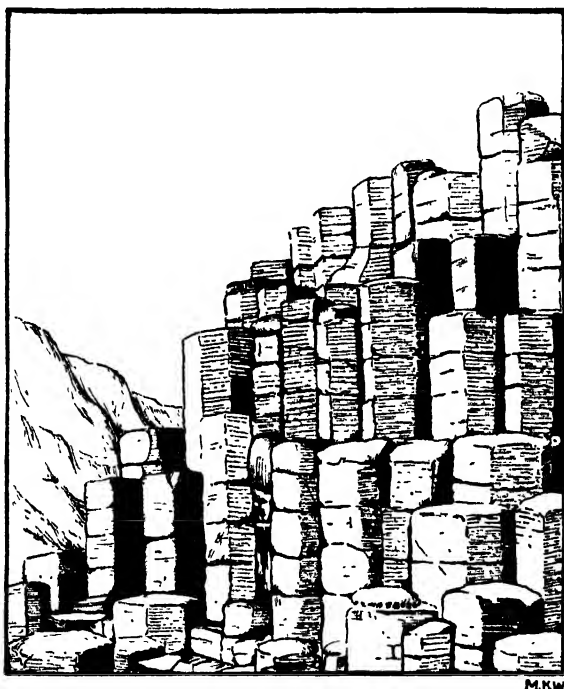


FIG. 58

Columnar basalt, Giant's Causeway, Co. Antrim, Ireland.

law, the columns are vertical in horizontal sills and flows; while in dykes they are horizontal if the walls are vertical. In many lavas, particularly those of Basic composition, three roughly parallel layers may be distinguished: an upper slaggy and vesicular portion; a central zone with somewhat irregular columns of small cross-section, and a basal layer, with more massive, regular, hexagonal columns. These latter have grown upwards from the slowly cooling base while the thin "wavy" columns grew downwards from the more rapidly cooling upper surface. Fine examples of columnar jointing are found in the Basic rocks of the Western Isles of Scotland, as at Staffa and in Skye, also at Elie and St. Andrews in Fifeshire.

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Occasionally the columns are subdivided into segments by a subsidiary cross-jointing, and usually such segmentation is accompanied by a spheroidal tendency, producing *ball-and-socket joints*. Spectacular examples are seen at the Giant's Causeway in Antrim, by far the finest columnar basalt in Britain¹ (Fig. 58).

MODES OF OCCURRENCE OF INTRUSIVE
IGNEOUS ROCKS

SHEET-INTRUSIONS

(a) **Dykes.**—In this category we include all intrusions of sheet-like form which are vertical or nearly so, at the time of intrusion. Dykes vary in thickness from a fraction of an inch to hundreds of feet, but the average width is probably between one and three feet.¹ As a consequence of their attitude the outcrops of dykes are not affected by the topography of the country in which they occur: they therefore appear as nearly straight lines on geological maps, maintaining a uniform direction, sometimes for long distances. Parallel suites of dykes or dyke-swarms were intruded when a portion of the earth's crust was subjected to tension. The Lorne dyke-swarm in South-West Scotland, with a characteristic north-east to south-west trend, and the Mull dyke-swarm are typical examples (see Figs. 59 and 148).

In other cases, however, a group of dykes has a radial disposition about a plutonic centre, as in the island of Rum. Here the tension resulted from the thrust of the magma against the roof (Fig. 148).

Usually the dykes in any one swarm are petrographically of the same type, and bear witness to the existence underground of extensive sources of (usually) basaltic magma. But the fact that the Cleveland dyke in the north of England, for example, can be traced almost continuously for 130 miles must not be taken to prove that the basin from which the magma was drawn necessarily extended over the same distance. Sometimes the internal structure of a dyke-rock suggests that the magma has been injected horizontally rather than vertically. Clearly, the fact that individual sheets, often only a foot or two in thickness, can be traced for such long distances,

¹ The largest rock-body to which the term "dyke" has been applied is the *misnamed* Great Dyke of Southern Rhodesia, one of the most remarkable basic-ultrabasic complexes in the world. Its sole dyke-like feature is its general form: it is some 3 miles across, but extends for 300 miles. There the resemblance to a dyke ends, for it is actually a narrow slice through a complex lopolith, or series of lopoliths, let down between parallel boundary faults and preserved in a graben.

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proves not only that the magma was exceptionally mobile, but also that it was injected rapidly.

Although as noted above the different members of a given dyke



FIG. 59

Sketch-map of South-West Highlands, showing a system of north-west basic dykes (of Tertiary age) following earlier and parallel lines of weakness.

(After H.M. Geol. Survey.)

MODES OF OCCURRENCE OF INTRUSIVE IGNEOUS ROCKS

swarm usually consist of the same rock-type, there are exceptions. The most outstanding example is probably a dyke swarm, of some 500 parallel dykes with constant strike, but varying from Acid to Ultrabasic in composition, occurring in South-West Madagascar.

Successive injections of the same type of magma into the same dyke-fissure form **multiple dykes**. On occasion the same fissure has been followed by two or more injections of contrasted types, giving **composite dykes**. The most general case is that in which an initial injection of basaltic magma appears to have been followed by a later injection of granitic material: the former is typically black and finely crystalline in the hand specimen, while the latter is often much lighter coloured, often red, and may carry relatively large crystals of quartz and feldspar. Examples of both multiple and composite dykes of Tertiary age occur in Arran¹ and Skye,² and are well displayed also in the north-easterly dyke-swarm of southern Jersey, Channel Islands.

Most petrologists agree that dykes are injected during periods of stress or tension in the earth's crust. The first stage in the emplacement of a dyke swarm must obviously be the development, in the area affected, of a series of parallel fractures, which admittedly might well be produced by shearing; but whether or not, the fractures must open to allow ingress of the basaltic material, and this is most readily visualized under tensional conditions operating at right angles to the given fissure. In the aggregate the "stretching" of the original belt of country is equal to the total thickness of all the dykes involved, and in some cases amounts to a most impressive figure.

A dyke which has been injected in the manner outlined above is said to be *dilational*. There is little doubt that all basaltic dykes are dilational, though evidence suggests that this is not always the case with granitic dykes. The critical evidence involves the relationship between the dyke and earlier structures in the country-rock. If such earlier structures are visibly displaced—by an amount depending upon the thickness of the dyke—where they cross the latter, the dyke is proved to be dilational. If, however, earlier structures are not displaced on crossing the dyke, the possibility of the latter having been formed partly or wholly by replacement along a fault must be taken into consideration.

(b) **Sills**.—The three-dimensional form of a sill is the same as that of a dyke; but the attitude is different. Sills are essentially *concordant* with the stratification of the country-rock, and therefore occur as sensibly horizontal sheets in untilted strata, though

¹ Gregory, J. W., and Tyrrell, G. W., *Proc. Geol. Assoc.*, **35** (1924), p. 413 and Pl. 26.

² Harker, A., "The Tertiary Igneous Rocks of Skye," *Mem. Geol. Surv.* (1904), pp. 201–7.

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subsequent tilting or folding may throw them into any attitude—including the vertical. It thus behaves at outcrop as if it were a part of the stratigraphical succession, and difficulty may be experienced in deciding whether a given sheet is sill or lava-flow. In some

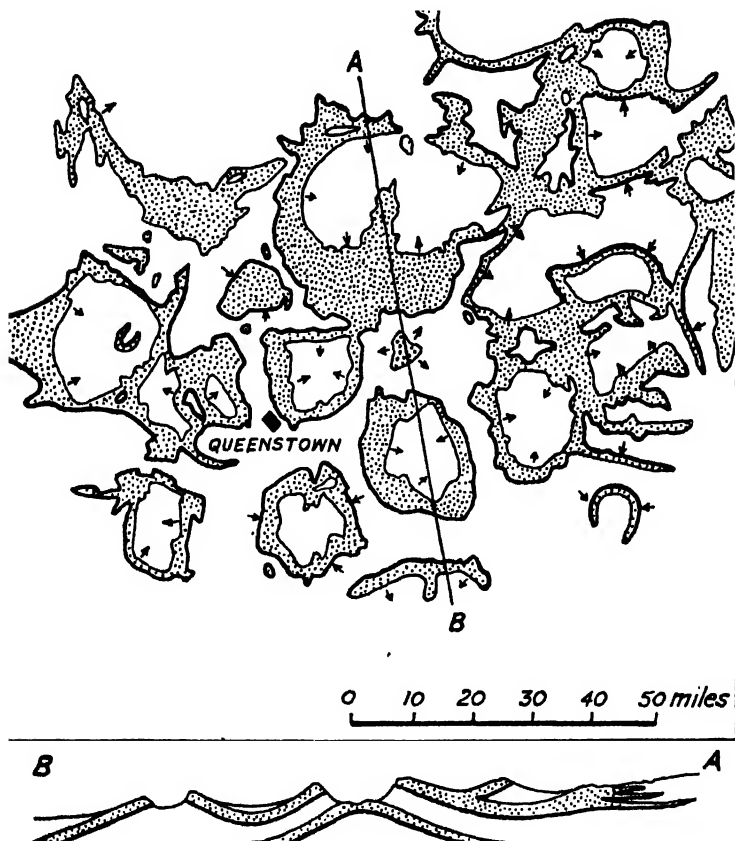


FIG. 60

Sketch map and section (vertical scale exaggerated) of thick Karroo dolerite sills intruded into sediments of the Beaufort Series, South Africa.

(After Rogers and du Toit.)

cases the mass-characters of the igneous rock may be sufficiently distinctive (see p. 134); otherwise the following observations may be helpful. Although sills may keep to one horizon for considerable distances, they do sometimes transgress suddenly to a higher or a lower horizon, in a manner which would be impossible for a lava-flow. Similarly vein-like offshoots, particularly into the rock *above*

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the igneous sheet, would indicate that the latter is a sill rather than a flow, though sometimes sills have been injected into mud on the sea-floor, and may simulate lava-flows very closely as regards such features as vesicularity and contact alteration of the adjoining sedimentary materials.

As with dykes, sills may be multiple and, in other instances, composite. Again they consist dominantly of consolidated basaltic magma, and sills occur extensively underlying the areas of vast basaltic outpourings, noted above. They are also noteworthy in the Karroo in South Africa, not only on account of their dominance of the characteristic scenery, but also of the extraordinary extent of some of the individual sills: some extend over areas of from 3,000 to 5,000 square miles. The dissected outcrop of one of these giant sheets is shown in Fig. 60. It should be noted that the remarkable undulations, causing basin- and dome-like structures are not characteristic of sills in general.

In Britain, the most extensive sheet is the Great Whin Sill, which underlies much of north-eastern England, and is well exposed in Teesdale where its outcrop forms the falls known as High Force. The most impressive sill known to the writers forms a thick, quasi-horizontal sheet of almost black dolerite intruded into bright red granite which forms spectacular cliffs in the north of Jersey, Channel Islands. Other better known sills include those of Tertiary age, consisting of "quartz-porphyry" (a granitic rock) and outcropping on the coast of southern Arran (the Dippin Sills) in western Scotland.

LENS-SHAPED INTRUSIONS

(a) **Laccoliths.**—These result when the intrusion of viscous magma along a plane of weakness has caused an uparching of the strata above the intrusive mass. Ideally the floor remains flat; and in the classic description of intrusions of this type, from the Henry Mountains, Utah,¹ it was suggested that the laccolith was fed by means of a central pipe. This is, however, purely hypothetical, and it is just as likely that the intrusion was fed by, and is a local modification of, a sill. The hydrostatic pressure of the magma is believed to have caused the uparching of the roof. Fundamentally a laccolith has the shape of a plano-convex lens, lying flat-side down.

In sills the lateral dimensions are very great compared with the thickness; but in the laccolithic type of intrusion, the latter varies between one-third and one-seventh of the diameter. Laccoliths may be multiple, as in the case of the so-called cedar-tree laccoliths.

(b) **Phacoliths.**—This name is applied to lens-shaped intrusions

¹ Gilbert, W. S., "The Geology of the Henry Mountains," *U.S. Geol. Surv.*, Washington, 1877.

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of concavo-convex shape situated in the core of an anticline. The phacolith owes its shape to intrusion into pre-existing folded strata, while the laccolith, in theory, *causes* the arching of the roof. The type phacolith forms the Corndon, a gabbroic or doleritic intrusion occurring in the Shelve district of Shropshire.¹

(c) **Lopoliths.**—Intrusions of this type are the converse of phacoliths, as they have the concavo-convex shape, but are convex downwards, that is, the central part of the intrusion has been warped downwards. Some of the greatest Basic intrusions in the world are of this type. The Sudbury (Ontario) lopolith and the Duluth (Minnesota) gabbroic complex are outstanding examples. The term lopolith was first applied to the Duluth intrusion,² which forms a crescent-shaped outcrop at Duluth, bounding the western tip of Lake Superior. The crescentic outcrop resulted from tilting of the lopolith; but no such tilting has affected the Bushveld Complex,



FIG. 61

Diagrammatic E.-W. section across the Bushveld complex, Transvaal. A red "granite" overlies the basic rocks (chiefly norite). Various members of the Transvaal system, into which the lopolith is intruded, are indicated as follows: *stippled*, Black Reef Quartzite; *cross ruled*, Great Dolomite; outcrops of Pretoria Series above intrusion, P P P. (*After Daly, simplified.*)

which is the largest Basic-Ultrabasic intrusion in the world.³ It extends east-to-west for 300 miles and occupies a surface area of *20,000 square miles within which an extraordinary variety of Basic and Ultrabasic rock-types are magnificently displayed (Fig. 61). The Complex is strikingly layered, and differential weathering has etched out the structure so that the harder rocks form strike-ridges approximately parallel to the boundary of the lopolith and also to the strike features formed by massive quartzites which dip beneath the floor of the intrusion. The thickness of the lopolith may be as much as 6 miles; but what happens beneath its centre is anybody's guess. There must have been a feeder, of course, but whether it was pipe-like or funnel-shaped does not seem to us to matter much: the main thing is that the Complex is there, and will provide material for petrological studies for a long time to come.

Some major intrusions of layered gabbros do, however, possess a funnel-like shape, both as regards the attitude of layering and

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the form of the intrusion. This applies, for example, to the Freetown Complex in Sierra Leone.¹ The dip of the layered structures becomes increasingly steep towards the centre of the intrusion, and it may be inferred from geophysical evidence and borehole data that the floor is similarly inclined. It may be noted that towards the outer margin of the Complex where the layers are more gently inclined, they form scarp features comparable with those of the Bushveld Complex. Although the Freetown Complex shares certain features with the great Basic lopoliths, it seems best to regard it as a funnel intrusion.

Funnel intrusions² are generally much smaller than the Freetown example, having circular outcrops a few miles in diameter. The type example is provided by the Cortlandt or Peekskill Complex of New York, in which there are three foci towards which the layered structures converge. Other intrusions of the same general type have been noted in New Hampshire, and for these it has been suggested that the magma was forcefully emplaced, making room for itself by pushing aside and doming the country rock. Some of the Tertiary gabbroic intrusions in Scotland may have been emplaced in a similar manner.

The most spectacular and the most completely studied intrusion known to have a funnel shape is that of Skaergaard, Eastern Greenland.³ The greater part of the complex is occupied by a layered series of gabbros, varying considerably in composition from below upwards. The significance of the variation in composition and of the layering displayed is discussed in due course. It is probable that space was made for the intrusion initially by volcanic activity of the explosive type. The diameter of the complex is comparable with that of many volcanic calderas.

So-called funnel intrusions thus vary widely in character: at one extreme they may be scarcely distinguishable from lopoliths of sheet-like form; while at the other, steep and discordant contacts suggest affinity with ring complexes described below.

RING COMPLEXES

Just as lavas can be referred to two main types of extrusion, the one of a widespread character (plateau lavas), and the other of a central and localized type (cone or central volcanoes), so with intrusive

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complexes. The dyke swarms and widespread sills are of the first kind and stand in strong contrast to the central type complexes, in which the full force of igneous activity was brought to a focus over a small area. The classical area for studying this type of complex is in the Inner Hebrides of Scotland, the chief individual centres occurring in Mull, Skye and Ardnamurchan. From the example in Mull, it is inferred that the intrusive complex represents the basal wreck of a large volcano of central type.

The crustal forces operating in central complexes result in the formation of characteristic crescentic intrusions. Only their essential characters are considered here, since examples are described in greater detail in the last section of this book.

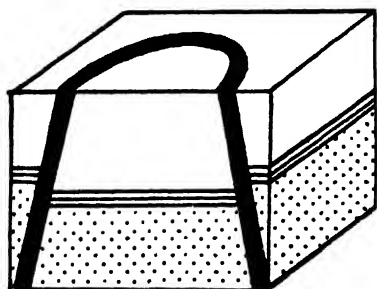


FIG. 62

Block-diagram of a single ring-dyke injected through horizontal strata.

(a) **Ring-dykes.**¹—This term covers intrusions which have arcuate outcrops, the radius of which seldom exceeds two or three miles. The inner and outer walls of the intrusion may be parallel at ground level, or they may have different degrees of curvature so that when traced along the strike they converge and eventually intersect, forming a crescentic outcrop. Very occasionally the dykes form complete rings as in the case of the Ossipee Mountains ring-dyke in New Hampshire.²

Ring-dyke walls are usually steeply inclined or may be vertical, though evidence is necessarily scanty. The Hebridean ring-dykes are believed to have steeply inclined, outward-dipping walls, and to have been emplaced by down-faulting of a central mass of rock enclosed within a ring-fault. Such down-faulting would provide a potential cavity into which magma could be drawn. This mechanism, first proposed by Clough, Maufe and Bailey for the Devonian ring-intrusion of Glencoe, Scotland, and elaborated particularly by

¹ Richey, J. E., "Tertiary Ring Structures in Britain," *Trans. Geol. Soc., Glasgow*, 19 (1931-32), p. 45.

² Billings, M. P., "Mechanics of Igneous Intrusion in New Hampshire," *Daly Volume, Amer. Journ. Sci.* (1945), p. 40, with references to literature.

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Richey as an explanation of the manner of intrusion of the many Tertiary ring-dykes in Scotland, neatly provides a solution to the space problem, particularly if it can be demonstrated that the walls dip outwards. However, there are instances for which alternative explanations must be sought. Some ring intrusions in the Hebridean Province have widths up to a mile, and furthermore possess internal flow-structures inconsistent with a simple hypothesis of subsidence; in some significant cases the walls locally at least dip inwards. It is possible that ring-faulting may be accompanied by the formation of a zone of fracturing and brecciation into which magma may be able to stope its way.

Ring-dykes may occur singly, but frequently they form **ring-complexes** consisting of several individual intrusions arranged concentrically about a common centre. The order of intrusion of the several members of the Complex may be inferred on the evidence of included fragments of an earlier member in a later one. Chilling of one member against another may occur but is apparently rare, presumably because the several members of the Complex were emplaced within a relatively short interval of time. Thin partitions or "screens" of country rock may separate neighbouring ring-dykes.

Although the rocks forming ring-dykes are often coarse-grained and therefore by definition fall in the "plutonic" category, there is no doubt that many, if not all ring-dyke complexes, were directly connected with surface vulcanicity, particularly with calderas. The near-surface character of some of the rocks involved is proved by the occurrence of such features as fine-grained texture, flow banding and explosive brecciation.¹

Theoretically any kind of igneous rock capable of intrusion may, on occasion, form a ring-dyke; but actually granitic rocks are far more commonly represented among such rock-bodies than those of other types. This may possibly be a consequence of the relatively low specific gravity of granitic magma. Among noteworthy examples of ring-complexes are those discovered and (in part) described within recent years in several different parts of Africa. Many of these include varieties of riebeckite-granites,² carbonatites and nepheline-bearing rocks. In Britain the triple ring Complex of the Ardnamurchan Peninsula in western Scotland is of special interest as among the rocks involved are Basic gabbros and "eucrites," Intermediate monzonites, and more granitic types including tonalite.

(b) **Cauldron Subsidence Intrusions; Bysmaliths.**—One further type

¹ Richey, J. E., "Association of explosion-brecciation and plutonic intrusion in the British Tertiary Igneous Province," *Bull. Volc.*, Série 11 (1940), p. 157.

² Jacobson, R. R. E., Macleod, W. M., and Black, R., "Ring-Complexes in the Younger Granite Province of Northern Nigeria," *Mem. Geol. Soc., London*, No. 1 (1958).

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of intrusion must be mentioned here, for it also is connected with a type of ring-faulting. As defined by Iddings,¹ a *bysmalith* is an injected body, having the shape of a cone or cylinder, which has either penetrated to the surface or terminates in a dome of strata like that over a *laccolith*. The "plutonic plug" of Russell² is a similar conception. Vertical displacement with faulting is the characteristic of this method of intrusion. Mount Holmes, in Yellowstone Park, is cited by Iddings as a type.

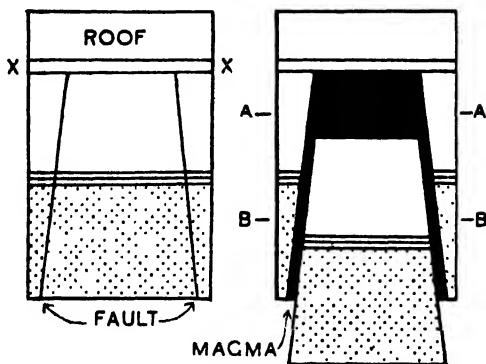


FIG. 63

Diagram sections to illustrate intrusion by cauldron subsidence.

It is clear that if the roughly cylindrical fracture of Fig. 63 had failed to reach the surface, or alternatively, if the sinking block had been terminated upwards by a horizontal plane of weakness (X in Fig. 63), a space would be formed *above* the sinking block and into it magma would be drawn. On consolidation this intrusion would be of cylindrical form, of circular cross section (as at level A in Fig. 63); it would have a flat floor, and at a certain level (as at B) it would give place to a ring-dyke. Such movements are referred to as "cauldron subsidences" and they are due to "piston faulting." The classical British example occurs at Glencoe in South-West Scotland, described by Bailey, Clough and Maufe. An intrusion may be inferred to have been emplaced in this manner if the following conditions are satisfied: (1) the environment is one in which there is a regime of ring-faulting; (2) the outcrop of the intrusion is approximately circular; (3) the diameter is of the order of 5 to 15 miles; (4) the contacts are sharp and steeply inclined outwards, though they may be vertical; (5) the surrounding rocks are undisturbed, or show a terminal curvature in the sense consistent with the sinking of the central cylindrical block. If, in addition, part

¹ *Journ. of Geol.*, 6 (1898), p. 707.

² *Ibid.*, 4 (1896), pp. 23-43.

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of the floor is exposed, or if there is a sharp turn-over from steep walls to flat roof, the evidence is much strengthened.

(c) **Cone-Sheets.**—The Tertiary intrusions of the Hebridean Province provide many examples of basaltic “dykes” aptly termed cone-sheets by E. B. Bailey. This three-dimensional term replaces A. Harker’s two-dimensional term, “inclined sheets.” Cone-sheets are normally only a few feet thick, and are almost invariably basaltic in composition. It is virtually impossible to prove that an

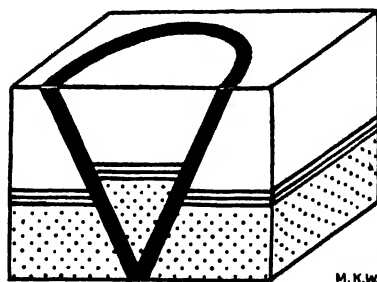


FIG. 64

Diagrammatic representation of a single cone-sheet, idealized to show the essential features of its shape, outcrop, and relationship to adjacent rocks. When comparing Figs. 63 and 64, remember the difference in relative dimensions of ring-dyke and cone-sheet. The converging walls of the latter doubtless reach the “magma reservoir” at some point *above* the apex of the cone.

individual cone-sheet has the ideal form shown in Fig. 64; but the diagram does embody the essential features of this type of intrusion: a curved outcrop at ground level, this being a segment of the circle shown in the diagram, and an inward dip towards a common focal point. In any area mapped in detail the extraordinary parallelism of the outcrops, their concentric arrangement about a centre and their constant dip are outstanding features. By extrapolation the dip observed at the surface may be extended downwards, and indicates the depth of the magma reservoir from which the material of the cone-sheet swarm was derived. Estimates relating to Hebridean centres range from 3 to 5 miles below the present surface.

The mechanism of intrusion of a cone-sheet is implicit in the diagram. The upper surface (“hanging wall”) is separated from the lower surface (“foot wall”) by the thickness of the dyke-rock, the intrusion of which must have involved an uplift which is measurable in favourable circumstances. If the cone-sheet were of the ideal form, injection would involve an upward displacement of the cone of country rock lying within its outcrop. This applies to all members of such a cone-sheet complex, and in the aggregate the total amount of uplift may be of the order of some thousands of feet. To achieve

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this the magma must have been capable of exerting a powerful upward thrust. The conditions are thus exactly the opposite to those operating in the case of ring-dyke injection, for which a deficit of magma-pressure is essential. An interesting feature of some of the Hebridean volcanic centres is an alternation of ring-dyke and cone-sheet injection.

Cone-sheet complexes do not appear to be common outside the Hebridean volcanic province; but relatively recently cone-sheets of unexpected material have been discovered, forming parts of complexes involving nepheline-bearing rocks and carbonatites. The latter in some cases form a central plug of crystalline carbonate which occurs also in associated cone-sheets. These are identical in form and attitude with the better-known basaltic examples, though the material of which they are composed is so strikingly different. The Alno Complex in Sweden and certain African complexes display these features. In these occurrences it is probable that the concentric conical fractures into which the carbonatites were injected resulted from explosive volcanic activity, consistent with the high concentration of volatiles necessary to produce this association of rock-types.

The large ring-dykes and granitic masses which have been intruded as a result of displacement following ring-faulting, form a transition between the intrusions described so far, in which the form and mechanism of intrusion are reasonably well understood, and the great subjacent intrusions, with steeply plunging contacts which continue downwards to unknown depths, and whose three-dimensional shapes and mode of origin have occasioned so much controversy. There is certainly scope for wide differences of opinion on this major problem of petrology, and for markedly different interpretations of the field evidence.

SUBJACENT PLUTONS

Subjacent plutons are major intrusions which have no visible floor. The "walls" are generally steeply inclined and within the limits of observation these rock-bodies tend to increase in size with depth. The rocks occurring in these plutons are mostly quartz-bearing types and can be classified as "granites" in the widest sense. The following description may therefore be regarded as relating to the various modes of occurrence of granitic rocks, in contrast to basic intrusives which more commonly occur as dykes, sills, funnels and lopoliths described above. The distinction is not, of course, exclusive.

(a) **Batholiths.**—This term was introduced by Suess to connote the major, deep-seated intrusive masses of very large size occurring typically in the great mountain ranges, and generally elongated

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parallel to their tectonic trend. The most spectacular batholiths occur in the western fold-mountain ranges of North America. The Coast Range batholith of Alaska and British Columbia extends along the strike for approximately 1,100 miles and varies in width between 80 and 120 miles; it thus occupies a surface area of some 110,000 square miles. The Sierra Nevada batholith measures 400 miles by 40 to 70 miles in width. These figures are impressive; but

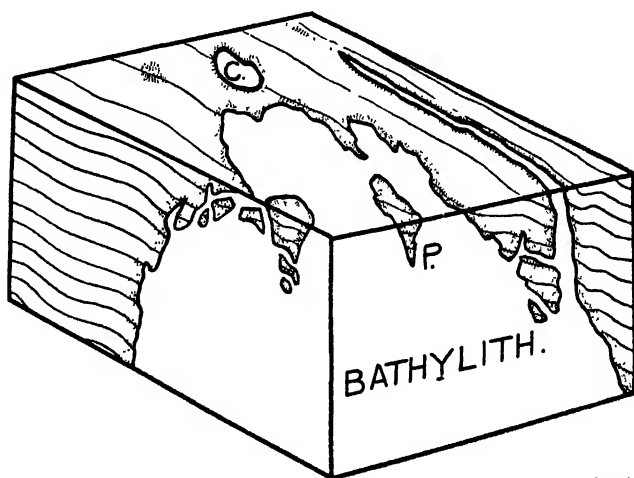


FIG. 65

Block-diagram showing the features of an ideal batholith (or bathylith). P. Roof pendant. C. Cupola. Metamorphic aureole stippled. (Modified from R. A. Daly.)

they are of great significance when we come to consider the problem of the mode (or modes) of emplacement of these rock-bodies.

Although very much smaller than the giant examples just cited, the Murrumbidgee batholithic complex, which occupies the southern tip of New South Wales, Australia, and extends to within a few miles of Canberra, is noteworthy as providing the most recent description of an intrusion of this calibre. Like the other great batholiths this Australian example is, in the real sense, a *complex* for it is built up of nine separate members as well as many minor ones, occupying in all an area of some 550 square miles. It would be pointless at this stage to enter into petrographic details, but it is clear that the several members of the complex were derived from one and the same source.¹

¹ Snelling, N. J., "The Geology and Petrology of the Murrumbidgee Batholith, Australia," *Q.J.G.S.*, 116 (1960), p. 187.

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The stripping off of the "roof" by denudation may, in the early stages, disclose isolated, irregular outcrops of igneous rock, which, apart from their petrographic similarity, convey no hint of the immensity of the pluton lying beneath. At a later stage, the isolated outcrops connect up, while the intervening stretches of highly metamorphosed roof-rock are reduced to small islands (roof-pendants). In addition to, or in place of a main outcrop, "cupolas" may rise from the hidden batholith, as satellitic intrusions.

In regions of great relief such as the Rocky Mountains and the Alps granite batholiths may be seen to maintain their characteristics to depths of the order of 10,000 to 15,000 feet. Further evidence of the downward extension of batholiths and other plutons has been obtained by geophysical means. Granitic rock has a low density compared with most other rocks, and this produces a deficiency of mass which can be easily detected by gravity surveying. It has been shown by this means that the granites of South-West England are connected underground and extend downwards for at least 10 miles.¹

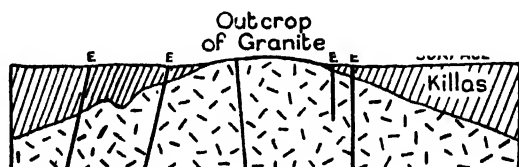


FIG. 66

Section through a granite dome intrusive in the killas of Cornwall and penetrated by dykes of elvan (E). (After MacAlister.)

(b) **Stocks** differ from batholiths only in size, and in some cases are merely cupolas exposed by the incomplete unroofing of a batholith. They may be defined as subjacent bodies less than 40 square miles in area. Like batholiths they have steeply plunging contacts and no visible floor.

(c) **Boss** is the term applied to stocks of circular cross-section. Some bosses and stocks have undoubtedly been emplaced by the cauldron-subsidence mechanism.

INTERNAL STRUCTURES IN SUBJACENT INTRUSIONS

Two structural elements are involved: firstly, the lineation of individual crystals and inclusions due to flow movements; and secondly, jointing due to fracturing after consolidation. Study of

¹ Bott, M. H. P. D., Masson-Smith, D., and Day, A. A., in *Phil. Trans. Roy. Soc., London*, 251 (1958), pp. 161-91.

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these phenomena was pioneered by H. Cloos¹ and the techniques have been summarized by R. Balk.²

Flow Structures develop as a consequence of the alignment of crystals and xenoliths during the act of intrusion. With regard to the first, the phenomenon is most clearly demonstrated by crystals of flattened, tabular habit, notably by feldspar phenocrysts. In this country the West of England granites, for example the "giant granite" of Dartmoor and Land's End, are typical in this respect. As regards xenoliths, these are often softened sufficiently to be drawn out into "schlieren" which appear as dark streaks in the normal granite. As a consequence of these flow structures, layers are formed which are normally parallel to the walls of the intrusion. In an intrusion with a domed roof these flow-layers or **platy flow-structures** dip outwards from the centre of the mass, and plunge steeply downwards in conformity with the dip of the actual surface of contact with the wall-rocks. They may thus be almost vertical in steep-walled stocks and bosses.

It is possible for the streaking out of these flow structures in the marginal zone of the intrusion to cause heterogeneous mineral banding similar to that commonly regarded as due to regional metamorphism. In the case of such primary gneissic banding, however, other metamorphic features, such as evidence of stress and cataclasis, should be absent.

It may be possible to detect the actual direction of flow of the magma by means of the parallel orientation of crystals of prismatic habit, such as hornblende, for example, or by the softening and elongation of xenoliths. These **linear flow structures** may be combined with layering, or they may occur independently; but in any case three-dimensional exposures are necessary to distinguish between the two types. It must be realized that these flow-structures are marginal phenomena; they tend to die out towards the centre of the mass, where the rock becomes massive, and even the feldspar phenocrysts, if present, become disorientated.

Joint Structures.—Once the outer crust of the mobile mass has solidified, no further flow is possible; but beneath the crust the magma may still exert a pressure on the crust and, by stretching it, produce tensional or "cross joints." Since the stretching follows the same direction as the linear flow of the previous plastic stage, the cross-joints are developed perpendicular to the linear flow direction of the latter.

There are, however, several sets of joints in a granite mass, and

¹ "Tectonik und Magma . . .," *Abh. Preuss. Geol. Landesanst.*, Heft 89, p. 1, and "Einführung in die tektonische Behandlung magmatischer Erscheinungen (Granittektonik)" (1925).

² "Structural Behaviour of Igneous Rocks," *Geol. Surv. America*, Mem. 5 (1937).

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enough has already been said to show that their proper interpretation is dependent upon the recognition of the flow structures. Some of the more important primary jointing systems are:

- (a) Cross joints, already described;
- (b) Longitudinal joints, striking parallel to the trend of flow lines, and probably developed as a result of weaknesses parallel to the aligned minerals;
- (c) Diagonal joints at angles of approximately 45° to the trend of the flow lines, formed as a result of the compression which operates at right angles to the flow.
- (d) Flat-lying joints, of which the origin is still obscure.

These joint directions may be followed by veins of aplite and pegmatite, or the joint planes may be coated with veneers of minerals of hydrothermal origin, thus indicating their primary origin. In the absence of such mineralization it may be impossible to distinguish

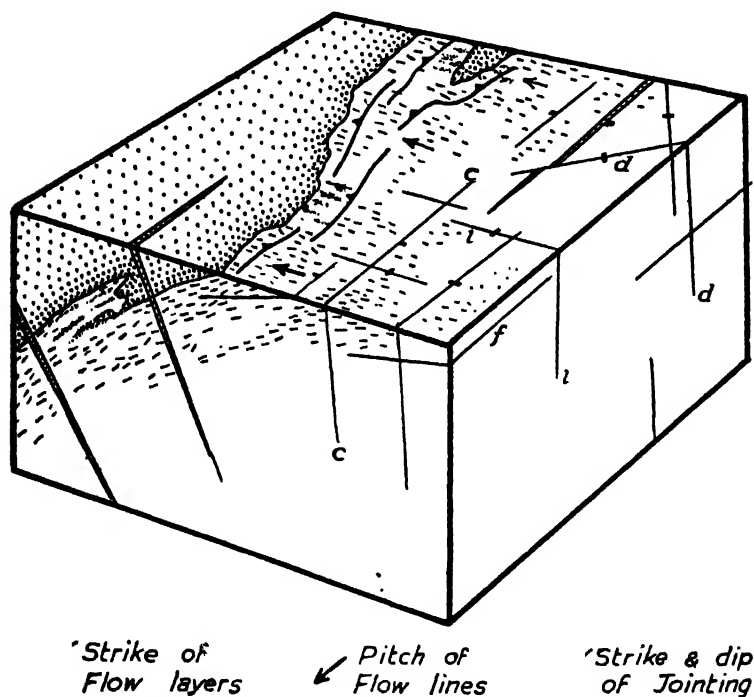


FIG. 67

Block diagram showing the relationship between flow structures and joint systems in an intrusive mass.

(c) cross joints; (d) diagonal joints; (f) flat-lying joints; (l) longitudinal joints.

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between primary joints connected with the act of intrusion, and those of later origin. Prominent among the latter are planes which develop parallel to the surface of the ground and named, rather misleadingly, exfoliation or bedding joints. In fact they may be highly perfect joint planes spaced widely apart, and occurring to a depth of many feet in the rock. Naturally only primary joints are of value for determining the possible shape of the rock-mass, and in practice most of such information is obtained from the study of flow structures combined with cross jointing alone.

Weathering takes place particularly easily along these various joint planes, so that exposed surfaces appear broken into angular blocks. On a freshly exposed surface it may be almost impossible to determine any directional properties in the rock, which appears massive and structureless. Nevertheless, planes of weakness, generally three almost rectangular sets, are normally present and can be detected by the experienced quarryman.

THE CONSOLIDATION OF MAGMA

THE original raw material of igneous rocks, or magma, may have a wide diversity of chemical and physical properties. Some of these have already been described when the nature of igneous rocks was discussed. For the purposes of the present chapter, however, it is assumed that a magma is originally a hot fluid which is essentially a melt of complex silicates and oxides, or more correctly, one capable of crystallizing out to yield these components. In addition, small amounts of ordinarily volatile components such as H_2O , Cl , CO_2 , etc., are present. The proportions of the various elements vary greatly, mainly according to whether the magma is highly siliceous (acid), or poor in silica (basic); but always the chief components include the following: Si , Al , Fe'' , Fe''' , Mg , Ca , Na , K , and oxygen. It so happens that in the silicates, the ratio of the other elements to oxygen is always such that their formulae may be written down in terms of oxides; for example, orthoclase $KAlSi_3O_8$ may be written as $K_2O \cdot Al_2O_3 \cdot 6SiO_2$. For this reason the convention has been adopted of stating the analyses of rocks in terms of the percentages of various oxides. It is well to bear in mind, however, that these "oxides" are not really present as such.

The cooling and crystallization of a melt of such complex composition is inevitably a very complicated process; but by careful study of the textural interrelationship between the minerals in rocks, it is possible to establish an order of crystallization. It is evident that the first minerals to crystallize are those which can be precipitated from a completely, or almost anhydrous melt, at high temperatures. These are the so-called **pyrogenetic minerals**, and include the majority of the silicates found as primary constituents in the basic rocks—olivines, most pyroxenes, the calcic plagioclases, etc. The separation of these pyrogenetic minerals leaves the liquid relatively enriched in H_2O and various other components of low atomic and molecular weights, which are known as the volatile, hyperfusible, or fugitive constituents. Several rock-forming minerals require for their formation a high concentration of these volatiles. These **hydatogenetic minerals** depend more on concentration of volatiles than on high temperature for their formation. Most of the

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alkali-rich minerals and those containing hydroxyl fall in this category.

The history of the cooling and crystallization of a magma can be divided into a number of stages, based largely on the dominance of the roles of temperatures and concentration of volatiles.

The **orthomagmatic stage** covers the separation of the pyrogenetic minerals, and in the case of a Basic rock, accounts for the crystallization of the greater part of the component minerals.

This is followed by a stage during which the portion still in the fluid condition has extremely low viscosity on account of the increasing concentration of volatiles, while the temperature is still fairly high, perhaps between 400° and 600° C. This leads to the development of crystals of exceptional size, distinctive of the **pegmatitic stage** of crystallization. It is possible that the more volatile fractions may sometimes separate out as independent gaseous phases under these conditions, especially if there is a local reduction of the external pressure within the system. Certain minerals, notably tourmaline, topaz and fluorite, are especially characteristic of such gaseous, or **pneumatolytic** conditions. The products of the pegmatitic stage are commonly segregated into veins and dykes, in which case they form highly distinctive rocks.

Finally the residual fluid may gradually pass towards the condition of low temperature aqueous solutions, and any deposition or replacement occurring at this stage is said to be **hydrothermal**.

The cooling history of a magma has been divided into more elaborate stages by some authors, who base their divisions on temperature limits. There is, of course, no direct way of determining the temperature at which any particular constituent has crystallized, but there are several mineral changes, for example the inversion of beta- to alpha-quartz, which are known to occur at fairly constant temperatures. The presence of these critical minerals, therefore, allows one to estimate approximate limits for the several stages.

These points on the "geological thermometer" are, however, too few and too much subject to variation under natural, as distinct from experimental conditions, to provide a wholly satisfactory basis on which to establish a complicated system of stages of crystallization.¹ In any case, the processes are actually continuous, overlapping of the stages is inevitable, and exact limits cannot in practice be assigned to them. In particular, it is often impossible to distinguish between phenomena produced during the pegmatitic-pneumatolytic stage and the hydrothermal stage. In some cases, indeed, it is possible only to differentiate between the products of

¹ A valuable summary and criticism of the many terms which have been so used is given by Shand, S. J.: "The Terminology of Late-Magmatic and Post-Magmatic Processes," *Journ. Geol.*, 52 (1944), p. 342.

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the orthomagmatic, primary crystallization on the one hand, and the late-stage modifications on the other. Some of the more important aspects of these two main, broad divisions are considered below.

PRIMARY, ORTHOMAGMATIC CRYSTALLIZATION

From an early date in the history of petrology much attention has been paid to the apparent order of crystallization of the minerals in rocks. It was observed that in a large number of cases there was a definite sequence. Applying the principle that if mineral "A" encloses or is moulded upon mineral "B" the latter must be of earlier formation than the former, the commonly observed sequence was found to be: accessory minerals, ferromagnesian minerals, feldspars, quartz.

However, many exceptions have been observed, and the principle is liable to misinterpretation since the observed order is that of completion, not commencement of crystallization. The so-called law depends on the premise that, since the initial composition of magmas varies only within quite narrow limits, there can be little variation in the order of crystallization.

In fact, however, the components in a complex silicate melt such as occurs in Nature mutually lower one another's freezing temperatures to a considerable degree. Further, the order of crystallization is dependent upon the concentration of a given component in the melt: no mineral will crystallize from a melt until the latter is effectively saturated with it, under the prevailing conditions.

An understanding of the laws governing the crystallization of minerals from a silicate melt has been made possible only by the experimental researches carried out at such specially equipped institutions as the Geophysical Laboratory, Washington. These experiments involve a high degree of specialist skill. Although the early experiments of this kind dealt only with the simplest silicate components at atmospheric pressures and under dry-melt conditions, the range and complexity of the experiments is being extended all the time, particularly in the field of high pressures involving volatile constituents, so that now even the most complex minerals are being synthesized, as the conditions of the experiments are brought ever closer to those of Nature.

The crystallization of a rock consisting of two components is instructive, and two possible cases will be considered: first, that in which the components are incapable of forming solid solutions (*i.e.*, they are immiscible in the solid state); and secondly, that in which the components are capable of forming a continuous series of solid solutions. In addition, the crystallization of simple three-component systems will be examined.

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The crystallization of a pair of minerals which do not form mixed crystals can be exemplified by means of a temperature-concentration diagram, in which the relative proportions of the two components are shown in percentages by abscissae, and the temperatures by ordinates. The case of an anorthite-diopside melt is illustrated in Fig. 68.

Starting with pure anorthite at the point A, the addition of an increasing proportion of diopside lowers the anorthite freezing-point as shown by the curve AE, and similarly the addition of

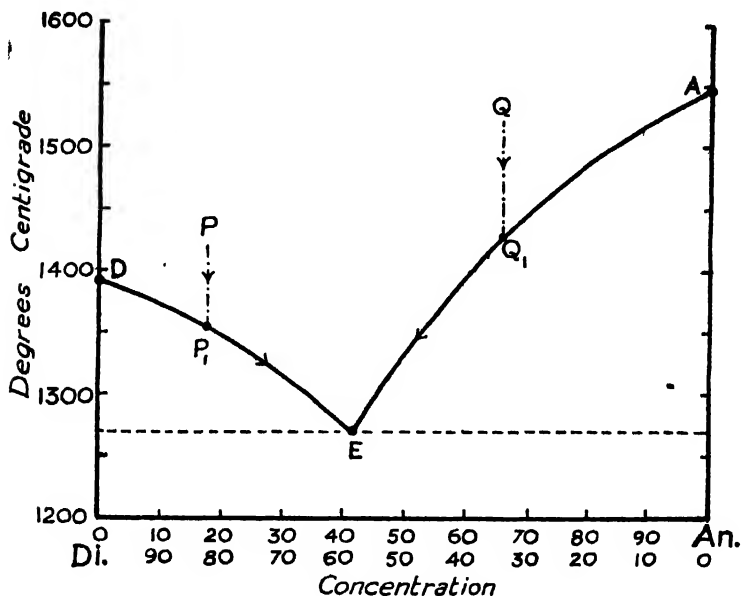


FIG. 68

Equilibrium conditions of the binary system diopside-anorthite.

(After N. L. Bowen.)

The explanation is given in the text.

anorthite to diopside produces the curve DE. The point at which the two curves intersect is the **eutectic point, E**.

Given a "melt" represented in composition and temperature by the point P, and subject to a falling temperature, no crystallization will take place until the temperature reaches P₁. At P₁ diopside (which is in excess of the eutectic proportion) begins to separate in the pure form. Its removal from the liquid causes the composition of the latter to change with falling temperatures towards the point E. On reaching this point diopside and anorthite crystallize together until all the liquid is exhausted. If, on the other hand, we start at a

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point Q, representing a liquid in which anorthite is in excess, a similar result is obtained, but pure anorthite first separates. It is to be noted that the solid formed by freezing at the eutectic point is a mechanical mixture of the two minerals, and not a solid solution. Thus it is seen that when a pair of minerals incapable of forming solid solutions (*mix-crystals*) are cooling together from a molten condition, there is a certain definite ratio of the one to the other in which they will crystallize out simultaneously in intimate admixture. Whichever mineral is in excess of this (the eutectic) proportion, will crystallize out first. It is important to realize (1) that the order of separation of the two components is independent of their freezing-points, and (2) that over a definite interval of time the two minerals are forming simultaneously.

The eutectic ratios for certain pairs of minerals have been established. Thus from an exhaustive study of gabbroic and noritic rocks Vogt has demonstrated that for hypersthene (or diallage) and labradorite the ratio is 35 : 65.¹ Investigations of the equilibrium conditions governing the crystallization of simple silicate melts have shown that for diopside and anorthite the ratio is 58 : 42, and for diopside and forsterite 88 : 12.²

An excellent illustration of the relation explained above is afforded by a case recorded by Harker from the Isle of Rum. In a series of rocks, consisting essentially of anorthite and olivine, the former is found to have crystallized out first when the rock is rich in that mineral, while in varieties rich in olivine the reverse holds good. Only when the two minerals occur in the eutectic ratio have they crystallized simultaneously.³

If in addition to the two minerals a third be present, say pyroxene in the case quoted, Nernst's law of the reduction of solubility between substances having a common ion,⁴ appears to govern the order of crystallization. Thus olivine and pyroxene have the ion (Mg,Fe) in common, consequently the solubility of the olivine is much reduced, and it invariably crystallizes before the feldspar, even when not present in very large amount.⁵

The crystallization of two minerals that form a continuous series of solid solutions is well illustrated by the plagioclase feldspars.⁶ In such a system, the solid and liquid in equilibrium at any temperature

¹ Vogt, J. H. L., "Physical Chemistry of the Crystallisation and Magmatic Differentiation of Igneous Rocks," *Journ. Geol.*, 29 (1921), p. 441.

² Bowen, N. L., "The System Diopside, Forsterite, Silica," *Amer. Journ. Sci.*, 38 (1914), p. 209.

³ Harker, A., "The Geology of the Small Isles of Inverness," *Mem. Geol. Surv.* (1908), p. 85.

⁴ Nernst, *Theoret. Chem.*, 3rd edn. (1904), p. 492.

⁵ Harker, A., "The Geology of the Small Isles of Inverness," *Mem. Geol. Surv.* (1908), p. 86.

⁶ Bowen, N. L., *Amer. Journ. Sci.*, 35 (1913), p. 583.

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are of different composition, the solid being always richer in the component with the higher freezing-point. In Fig 69, the curve marked "solidus" traces the change of composition of the solid with falling temperature, while that marked "liquidus" shows the corresponding change in composition of the liquid. Temperatures are plotted vertically, while the compositions in percentages of anorthite (An) and albite (Ab) are measured on the horizontal "composition line."¹ The course of crystallization is as follows: Starting with a melt of composition $Ab_{50}An_{50}$ at a temperature T_1 , no solid will be deposited until the point (a) is reached. At this temperature (T_2) a small amount of solid of composition (b) will be deposited. This solid (b) is in equilibrium with liquid (a) at temperature T_2 (and similarly for every other pair of points on the two curves cut by horizontal lines representing successive temperature changes). The separation of anorthite-rich solid has, however, altered the composition of the liquid phase which is richer in albite than originally. Let us suppose that the fall of temperature is temporarily arrested at T_3 , when solid (c) is in equilibrium with liquid (d). The earlier formed solid (b) will react with this liquid, the excess of anorthite will be leached out, and equilibrium will be restored when all the solid has the composition (c). But the fall of temperature is continuous, the change in composition of the liquid is continuous, so the solid also gradually changes in composition, approaching closer and closer to $Ab_{50}An_{50}$. The whole mass will have this composition at the temperature of final consolidation, T_4 .

During the cooling of such a system, only a single liquid phase can exist; while if equilibrium is fully established, then there is also only one homogeneous solid phase. The difference of temperature between the commencement and completion of the act of crystallization is called the "crystallization interval" (T_2-T_4). An interesting corollary is that the freezing-point and melting-point of such a system are not the same: freezing begins at T_2 while melting begins at T_4 .

Hitherto we have assumed that cooling was taking place very slowly and that equilibrium was fully established at every stage. Such, however, is not invariably the case. It frequently happens that cooling is so rapid that, on account of the high viscosity of the melt, equilibrium between solid and liquid is not fully established, and hence the growing crystals vary in composition, and in optical and other physical characters from the centre outwards. Each successive layer or zone deposited on the nucleus retains its original

¹ The composition of the solids and liquids represented by the points (a)-(d) are found by dropping perpendiculars from the point under consideration to the composition line: thus (c) has the composition $Ab_{30}An_{70}$.

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character: there is no time for reaction with the liquid portion of the system to take place. In this way **zoned crystals** are formed, and are specially common in the plagioclase and pyroxene groups. In the former case the nucleus of a crystal having the bulk composition of

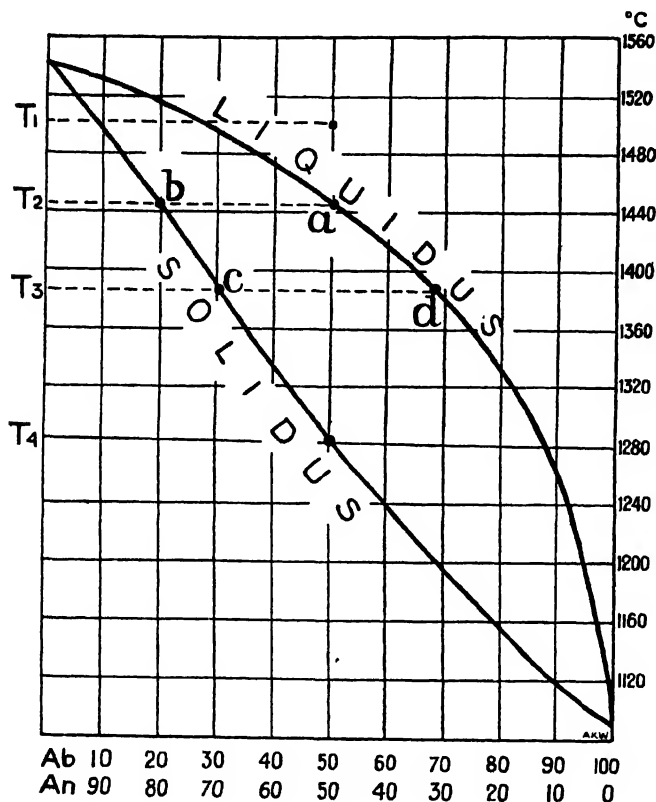


FIG. 69

Diagram to illustrate the crystallization of the plagioclase feldspars.
(After N. L. Bowen.)

(From *Amer. Journ. Sci.*, 35, 1913.)

labradorite may be almost pure anorthite, $\text{Ab}_{20}\text{An}_{80}$ (*b*), while the last drop of liquid, *i.e.* the last zone added to the growing crystals, is nearly pure albite, $\text{Ab}_{88}\text{An}_{12}$ (Fig. 69).

Although we have considered the plagioclases only, the importance of the principles illustrated by this one example may be gauged from the fact that with the exception of quartz, all the important rock-forming silicates are members of similar solid solution series

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and their crystallization must follow a similar course: the comparable diagram for the olivine series is illustrated in Fig. 4.

The course of crystallization is naturally more complicated when dealing with two components, both of variable composition, and we choose the pyroxenes to illustrate the principles involved. It was pointed out in the section on pyroxenes (p. 46) that during cooling of a melt the compositions change progressively as the temperature falls, the changes in general involving iron-enrichment. Further, in normal basic magmas over the higher temperature range, two series of pyroxenes of contrasted compositions crystallize simultaneously. When the compositions achieve a certain degree of iron-enrichment, however, the two pyroxenes give place to a single pyroxene-phase, corresponding in composition with ferroaugite.

To understand more fully the course of crystallization at any point along the fractionation series it is necessary to consider a cross section at right angles to the latter (as shown in Fig. 10), through the mid-point of the En-Fs join. It lies well within the two-pyroxene field mentioned above. The composition of the melt may lie anywhere within the system, of course, but arbitrarily it is assumed to be represented by the point *O* in Fig. 70. Cooling of the melt to a point on the liquidus phase boundary causes an augite to crystallize. Its composition is obtained by tracing along an isotherm (the horizontal broken line) to t_1 , then dropping a perpendicular from *q* on the solidus curve to the base line, where the composition corresponds with a_1 . This is the first augite to crystallize from a melt of composition *O*, and the crystal *q* is in equilibrium with the liquid *p* at temperature t_1 . Removal of this early augite enriches the melt in MgFe; and as cooling proceeds the composition changes until the point *r* is reached. This corresponds with the eutectic point of the previous diagram (Fig. 68); but we are now dealing with two isomorphous components and the curved lines meeting in *r* represent a section across a valley, the "long profile" of which extends at right angles to the diagram. The line of the valley may be referred to as the *cotectic* line.

At the temperature t_2 two pyroxenes crystallize simultaneously, augite of composition a_2 being joined by pigeonite. No further fall of temperature occurs until all the liquid is used up. Such further changes as do occur involve the crystals already formed and may be referred to as sub-solidus changes. Inversions and exsolution phenomena come into this category. The space below the augite and pigeonite solidus curves and between the augite and hypersthene field boundaries constitutes an immiscibility gap. The field boundaries referred to are called solvus curves, and it will be noted that they diverge downwards. At the temperature t_3 augite (*v*), more calcic than a_2 , is in equilibrium with pigeonite (*u*) richer in MgFe than that which

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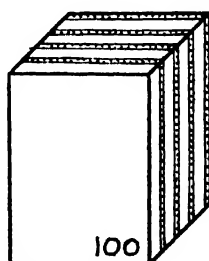
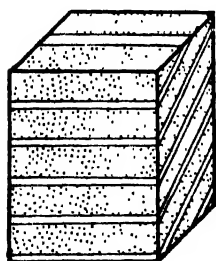
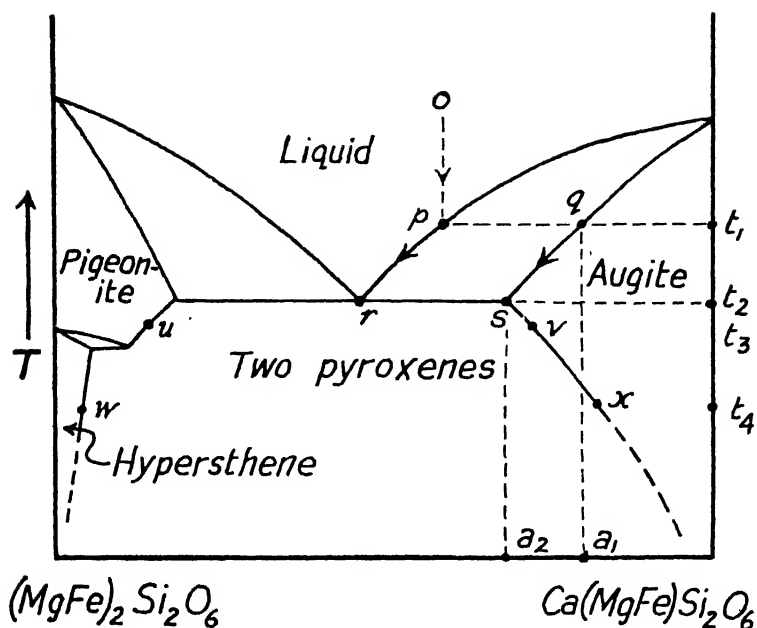


FIG. 70

Diagram illustrating the course of crystallization of pyroxenes; and (below) rectangular block-diagrams showing the orientation of exsolution lamellae. On the left are shown lamellae of augite parallel to the Monoclinic (001) planes, with orthopyroxene as host mineral. On the right, lamellae of orthopyroxene parallel to (100) are enclosed in augite.

(After H. H. Hess and A. Poldervaart.)

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first formed. At temperature t_4 still more calcic augite (x) is in equilibrium with hypersthene. These changes in composition as cooling proceeds are, of course, consistent with the tendency of both minerals to exsolve as explained under "pyroxenes" in the mineral section of the book.

The order of crystallization in a system of three components can be easily treated along similar lines. For the sake of simplicity we will assume that the components A, B and C do not form solid solutions. Each pair of components can form a binary eutectic, and there is in addition a ternary eutectic of three components ABC.

A system of this kind can be represented by a **triangular diagram** which is the projection of a solid model on the plane of its base.

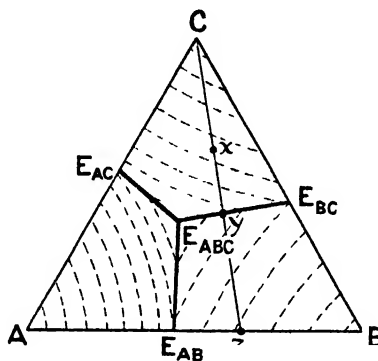


FIG. 71

Triangular diagram to illustrate the course of crystallization in a three-component system.

Each of the pure components is placed at a corner of the triangle, and each of the sides of the latter is divided into parts to represent percentages. Each point on the sides of the triangle then represents a mixture of two components; while every point inside the triangle represents a mixture of all three components. The ternary eutectic is one such point, though this does not necessarily lie at or even near the centre.

From each of these points representing compositions, perpendiculars are drawn proportional in length to the freezing-point of the mixture. It is obvious that there is an infinite number of such mixtures and that the perpendiculars would make a solid model rather like a trigonal prism with an irregular upper surface. The shape of this surface is indicated on the triangular diagram by drawing temperature-contours on it and then projecting them on to the base, as in the preparation of an ordinary contour map (Fig. 71).

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It will be noted by those familiar with contouring that the topography represented by this "map" comprises three slightly curved surfaces meeting in three valleys (shown by heavy black lines) converging on the lowest point E. The valley floors are phase boundaries which divide the whole triangle into three fields of existence for A, B and C respectively, called the A-field, etc. Any point in the triangle represents a particular composition and a particular temperature. The course of crystallization of a mixture represented by the point x will be as follows: since the mixture clearly contains the component C in excess of the binary eutectic ratio BC, this excess begins to separate out, and the composition of the mixture changes along the line joining C to x and continued until it strikes the cotectic-line E_{BC} E_{ABC} , at y . The solution is then also saturated for the component B, and B and C separate together as the binary eutectic. The composition then changes along the line y - E_{ABC} , till finally the ternary eutectic is reached at E_{ABC} . For any other point the course of crystallization is similar, except that, if the original mixture lies within the B-field, B crystallizes first; if it lies within the A-field then A begins to crystallize first. A line such as Cxyz in Fig. 71 illustrates an important principle which is useful in the consideration of such diagrams, namely that every point on it represents a constant ratio of A to B, the proportion of C only varying (from 0 per cent at z to 100 per cent at C).

When a solid model is made on the principle outlined above, it is readily seen that each of the vertical planes standing on the lines AB, BC and CA is a binary eutectic diagram like Fig. 68.

The outstanding points of interest are (1) that the order of separation of the minerals is determined chiefly by their relative concentration in the mixture, and (2) that the periods of crystallization of the several components overlap: during a definite period two components, and at a slightly later stage three components, form simultaneously.

The theoretical example of a three-component system just described can be regarded as an aid to understanding the more complex system described below: this involves the silica minerals, feldspars and feldspathoids (Fig. 72). The system is of vital interest in petrology since it covers the field of composition of alkali granites and also has an important bearing on the origin of feldspathoidal rocks. The continuous lines in all the diagrams correspond with phase-boundaries. In A the boundary surfaces between melts and the several minerals is contoured with broken lines, the figures indicating hundreds of degrees Centigrade. Under dry-melt, experimental conditions the minerals formed include high-temperature forms such as carnegite (car.), cristobalite (cr.), and tridymite (tr.). These are metastable under conditions of slow cooling and are

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converted into their more familiar low-temperature equivalents, nepheline and quartz.

The relatively large field of stability of leucite at high temperatures should be noted. A contoured diagram of this kind indicates the

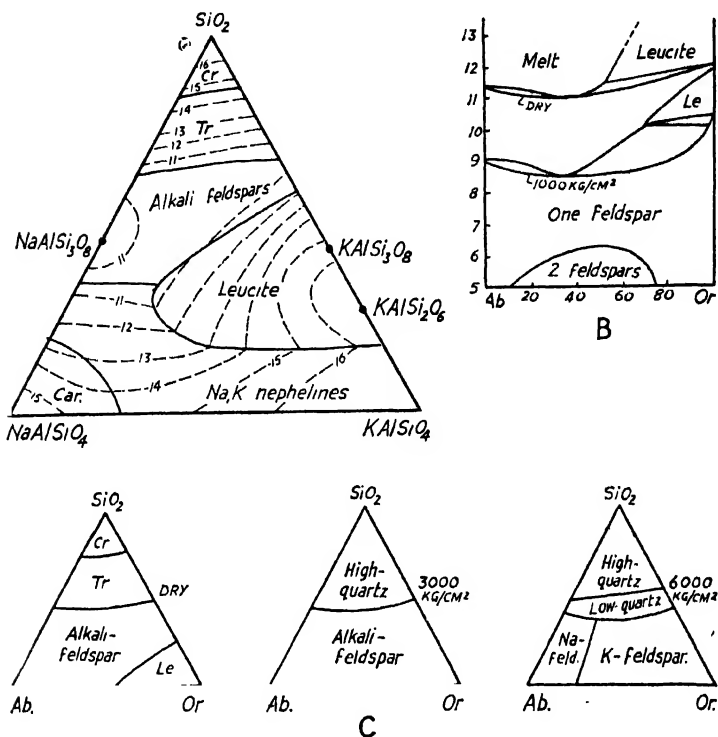


FIG. 72

Phase equilibrium diagrams relating to the occurrence of quartz, the alkali-feldspars and feldspathoids.

(A after Schairer and Bowen; B and C after Tuttle and Bowen.)

For explanation see text.

nature of the crystals first precipitated from a melt of any composition covered by the triangle. Separation of the crystals causes the composition of the melt to change, the change always following a "downhill" route on the contoured surface. The ultimate destination is one of the thermal "valleys" following the cotectic boundaries between either quartz and the alkali-feldspars, or the latter and nepheline.

Fig. 72B may be regarded as a vertical section along the join between albite and orthoclase. The diagram also shows the solidus

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curves, and the effects of reaction can be assessed between the crystals first precipitated and the liquid. For example, leucite precipitated from a potassic melt of appropriate composition reacts with the liquid at a lower temperature, being converted into orthoclase, if cooling is sufficiently slow.

The relations appertaining to dry-melt conditions should be compared with those relating to an aqueous melt. Even if the water-vapour pressure is only moderate (equivalent to 1,000 kg/cms²) the melting temperatures are lowered by hundreds of degrees. Another noteworthy effect is the reduction of the leucite field. Ultimately leucite disappears altogether if the water content is sufficiently high. This is the main reason why leucite is absent from plutonic rocks, regardless of whether the latter are saturated with silica or not.

Fig 72B also shows the sub-solidus (or so-called solvus) curve of unmixing of the synthetic alkali-feldspars.¹ A high-temperature homogeneous feldspar, for example sanidine, will unmix under conditions of sufficiently slow cooling to form various kinds of perthitic or antiperthitic intergrowths, the compositions of the sodic and potassic phases at any given temperature below 660° C. being indicated by the solvus curve. Only the top of the curve is shown in the diagram; but the divergence continues for temperatures below those indicated, so that efficient unmixing, seen in the most slowly cooled rocks, results in a relatively "pure" K'-feldspar host containing lamellae of Na-feldspar or vice versa.

Fig. 72C shows the phases that occur in the system SiO₂-albite-orthoclase under various conditions of water-vapour pressure. The first triangle, showing the dry-melt conditions, is merely a duplication of the upper part of A, inserted for comparison. The second shows the effect of lowering the liquidus temperatures with sufficient water-content to eliminate leucite and to produce β -quartz. With very high water-vapour pressure the liquidus surface is so far depressed that it intersects the solvus and therefore two alkali-feldspars are simultaneously precipitated from the melt.

Since the pioneer researches of N. L. Bowen, much emphasis has been placed upon the **reaction principle** in rock-building, that is on the mutual interaction which takes place in a cooling magma between the solid and liquid phases. This reaction may be continuous, giving homogeneous solid solutions, as in plagioclases, or it may be discontinuous, taking place only at definite temperature intervals. The latter type is illustrated by the replacement, completely or in part, of olivine by pyroxene, and of pyroxene by primary

¹ The solvus curve for the alkali feldspars may be compared with that of the pyroxenes shown in Fig. 70: the main difference lies in the fact that the pyroxene solvus is "beheaded" by intersection with the solidus curves.

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hornblende. Such a series of minerals, when arranged in the order in which the transformations occur, constitutes a *discontinuous reaction series*. Bowen has drawn attention to the following series: olivine \rightarrow orthorhombic pyroxene \rightarrow common augite \rightarrow hornblende \rightarrow biotite; leucite \rightarrow orthoclase; anorthite \rightarrow albite; common augite \rightarrow sodic augite; common hornblende \rightarrow sodic hornblende \rightarrow lepidomelane.

It should be realized that almost all the rock-forming silicates are members of isomorphous groups, each one of which is a continuous reaction series; that each of these in turn participates in the changes hinted at above. Further, taking the simplest possible view of the course of crystallization in a natural magma, it has been established that there are at least two series of changes taking place concurrently, and to some extent overlapping and interdependent. The one involves the coloured silicates (the mafic constituents) and the other the felsic components. N. L. Bowen¹ represents them thus:

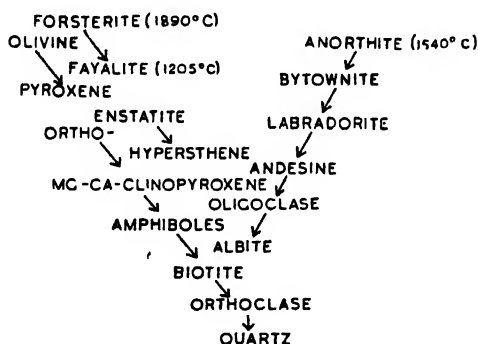


Table showing discontinuous reaction series (left-hand side) and continuous reaction series of the plagioclase feldspars (right-hand side). (After N. L. Bowen.)

In the discontinuous series represented on the left-hand side of the table above, apart from accessories the first silicate minerals to appear are the olivines, and of these, magnesium-rich precede iron-rich types.² Their separation from the magma takes place at high temperatures.

At a certain lower temperature these react with the magma and tend to be "made over" into magnesium-rich pyroxenes. These again are stable only over a limited temperature range and in turn tend

¹ *The Evolution of the Igneous Rocks* (1928), p. 60.

² Bowen, N. L., and Schairer, J. F., "The System MgO-FeO-SiO₂," *Amer. Journ. Sci.*, 29 (1935), p. 151.

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to give place to monoclinic pyroxenes containing the Ca-bearing diopside molecule. So the changes suggested by the table continue.

Meanwhile at high temperatures calcic plagioclase will crystallize, and with falling temperature will react with the magma, changing its composition steadily towards the sodic pole—the albite end of the continuous series. The interrelation of the two series of changes is suggested by the fact that the conversion anorthite \rightarrow albite releases CaAl_2 , while the change in the pyroxenes involves the addition of CaAl_2 .

It may be noted that there is a continual increase in complexity of atomic structure in passing from olivine to biotite. In olivine the structure is as simple as possible, the crystals being constructed of separated SiO_4 -tetrahedra. In the pyroxenes they are linked into chains; in the amphiboles they form bands; and in the micas they form extended sheets. The successive conversions involve abrupt changes in crystal atomic structure: each change is a passage from simple to complex.¹

THE LATER STAGES OF CRYSTALLIZATION: DEUTERIC PHENOMENA

Once we get down to the amphiboles and micas in Bowen's reaction series, the complexity of composition has until recently made experimental work almost impossible. This is true of all the hydroxyl-bearing silicates, and to those which have crystallized from a fluid enriched in volatiles, or fugitive constituents. The very term "fugitive" implies, as Shand meant it to do, that these constituents do not, in many cases, remain as part of the ultimate rock body. They may take part in several reactions and make others possible, and yet leave no trace of their passage. Thus it is necessary to face the task of interpreting an important part of the crystallization of an eruptive rock without knowledge of either the quality, the quantity or the ultimate whereabouts of the agents involved.

The important role of the fugitive constituents can be judged from the following facts.

1. When volatile-rich magma is erupted at the surface as lava, the fugitives are enabled to escape rapidly, and in so doing they immediately increase the viscosity of the lava, which solidifies rapidly, either as glass or as a cryptocrystalline aggregate. This rapid crystallization, coupled with the violent oxidation of escaping gases, accounts for the rise in temperature, amounting in some cases to a hundred or two degrees Centigrade, which is almost certainly

¹ Brammell, A., "Mineral Transformations and their Equations," *Science Progress*, No. 120, 1936.

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responsible for the corrosion of previously formed phenocrysts. The significance of the vesicular and amygdaloidal structure of certain lavas, and even minor intrusives that consolidated near the surface, is too obvious, in this connection, to need stressing.

2. In some active volcanic regions there exist powerful reminders of the activities of fugitive constituents in the form of **fumaroles**, which give off intermittent or continuous streams of gases, some of which form mineral deposits by sublimation. These volatiles also cause considerable alteration of the rocks surrounding the vent.

From this visible evidence provided by volcanic activity, we turn to consider the comparable effects produced in a deep-seated environment. Here direct observation is impossible, but it is reasonable to infer the probable course of events. One of the chief difficulties is to decide whether the active solutions arise entirely as a residuum left over after the completion of orthomagmatic crystallization, or whether they have been derived from some extraneous source. When the active solutions are directly of magmatic origin, all changes in mineral composition or in texture produced by them are termed **deuteric**. This term was introduced by J. J. Sederholm,¹ and may cover crystallization or alteration phenomena appropriate to both the pegmatitic and the hydrothermal stages. In the second case the active solutions are derived usually from a later intrusion, and any modification they effect is distinguished as **metasomatic**. Not infrequently it is impossible to distinguish between deuteric and metasomatic phenomena. We may thus be guilty of extending Sederholm's most useful term beyond its original meaning. The very acid and the highly alkaline rocks are usually products of magmas rich in H₂O and other fugitive constituents, and therefore tend to exhibit deuteric phenomena on an extensive scale, and in a variety of ways.

Since deuteric phenomena arise when the rock is very nearly solidified, it is natural that they should consist to a large extent of veining and replacement of earlier formed minerals. Several of these late-stage replacement phenomena are sufficiently distinctive and important to have names of their own: albitization, analcitization and chloritization are among the more important examples, while tourmalinization and silicification might also be considered in this category on occasion.

The term **albitization** covers a wide range of phenomena, though in essence, of course, it is simply the partial or complete replacement of earlier formed plagioclase or potassic feldspar by albite, which is stable in the presence of volatile-rich, lower temperature residual

¹ "On Syntactic Minerals and Related Phenomena," *Bull. Comm. Géol. Finlande*, No. 48 (1916), p. 134.

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solutions. At this point we consider only those aspects of albitization which may legitimately be regarded as deuteric: we omit for the time being the widespread development of albite-rich rocks on a regional scale, exemplified by the occurrence of the spilitic suite; and similarly the conversion of basic plagioclase into albite charged with inclusions of lime-rich minerals, due to low-grade metamorphism of the so-called regional type. Deuteric albitization, on the other hand, involves, at an early stage, the development of the less regular types of perthitic structure. The "patch-" and "injection-perthites" are presumably formed in this way, by the action of residual solutions on orthoclase or microcline. In slightly different circumstances the albite forms water-clear crystals interstitial to the minerals of earlier formation, or it may be deposited as a mantle around earlier feldspars. If the latter is orthoclase or microcline, there will be a striking contrast in appearance between the core and the external zone of the crystals, and this may be obvious in hand-specimens. If the core consists of plagioclase, however, there may be an almost imperceptible gradation towards the outer albite rim. The result may be a zoned crystal, indistinguishable from one produced by rapid cooling of an anhydrous melt. Thus there is no hard-and-fast dividing line between orthomagmatic and late-stage crystallizations: it is often impossible to say where one ends and the other begins. In fact, the distinction between the two stages only becomes apparent if the deuteric minerals are of a different nature from those of earlier formation, and if the former demonstrably replace the latter.

Analcite behaves in much the same manner as albite in many cases, especially as regards veining and progressive replacement of feldspars. But, like all late-stage minerals, it need not necessarily be replacive, but may crystallize in the interstices between crystals of earlier formation. When this is the case, although the crystal boundaries may be plane, with no sign of corrosion or embayment, there is clear evidence of the chemical activity of the residual solutions from which the analcite crystallized, particularly when the surrounding crystals are pyroxenes. For example in teschenites the pyroxene is a typical lilac-coloured titanaugite except where it is in contact with such "pockets" of analcite: here it is rimmed with bright green aegirine-augite. Thus analcite plays a dual role. That which occupies these interstitial areas must be regarded as the last of the primary minerals to crystallize out; but that which so clearly veins and replaces earlier feldspar crystals is just as definitely secondary by definition, although the division is arbitrary and artificial.

This double role is also played by **chlorite**. The several members of the chlorite group may replace and pseudomorph pyroxenes,

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amphiboles and micas, generally with the development of a fibrous, and occasionally a spherulitic, habit. In certain rocks, however, pellucid grains of chlorite occur, bounded by crystal faces of the very minerals which elsewhere are replaced by it. Obviously such interstitial chlorite is primary—it has replaced nothing. Once again we are seeing the result of overlap between the final primary crystallization and deuteric replacement.

Because of their late-stage origin, there is a widespread tendency to call such minerals as chlorite and analcite secondary, whatever their mode of origin. The only logical use of this term, however, is to restrict it to minerals which demonstrably replace others of earlier formation. If one of these late-stage minerals has an interstitial mode of occurrence, there is no alternative to calling it primary unless, of course, there are actual relics of an earlier mineral visible.

Two particularly interesting and much debated phenomena remain for consideration: they are (1) interstitial micrographic intergrowth of alkali-feldspar and quartz occurring in certain basic rocks; and (2) the peculiar quartz-plagioclase intergrowth, known as myrmekite, which is relatively common in acid and intermediate rocks.

In regard to the first, various hypotheses have been suggested to explain the intergrowth, which is commonly called **micropegmatite**. Vogt¹ claimed that micropegmatite results from the simultaneous crystallization of orthoclase and quartz in eutectic proportions in residual solutions. Other workers have found evidence that similar structures arise from the introduction of alkalis and lime into highly siliceous rocks; that is, micropegmatite may result from the feldspathization of quartz.² The converse process—silicification of feldspar—has also been suggested as a further possibility.

Both silicification of feldspar and feldspathization of quartz undoubtedly form types of quartz-feldspar intergrowth, though it is doubtful if the resulting structures are as geometrically perfect as those resulting from simultaneous crystallization. In the case of interstitial micropegmatite, however, any replacement hypotheses appear to be untenable, since the intergrowths are moulded upon, and radiate from, unaltered and euhedral feldspar crystals. In other words, the micropegmatite crystallizes, without any replacement, from an interstitial residuum.

Myrmekite is quite different, both in composition and in mode of occurrence. It consists of lobate patches, often described as cauliflower-like in form, of plagioclase riddled with small “vermicules” of quartz. The lobes normally have grown on plagioclase

¹ “The Physical Chemistry of Crystallization and Magmatic Differentiation of Igneous Rocks,” *Journ. Geol.*, 31 (1923), p. 245.

² Reynolds, D. L., “Demonstrations in Petrogenesis from Kiloran Bay, Colonsay,” *Min. Mag.*, 24 (1936), p. 367.

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into microcline (Fig. 73). F. Becke¹ has estimated that the amount of quartz in the intergrowth increases as the host plagioclase becomes more basic. This fits in well with the hypothesis that the quartz is released as a result of the replacement of potassic feldspar (the formula of which may in this connection be represented by $K_2O \cdot Al_2O_3 \cdot 6SiO_2$), by plagioclase which contains the anorthite component, represented by $CaO \cdot Al_2O_3 \cdot 2SiO_2$.

Besides releasing silica, this reaction would also release potassium

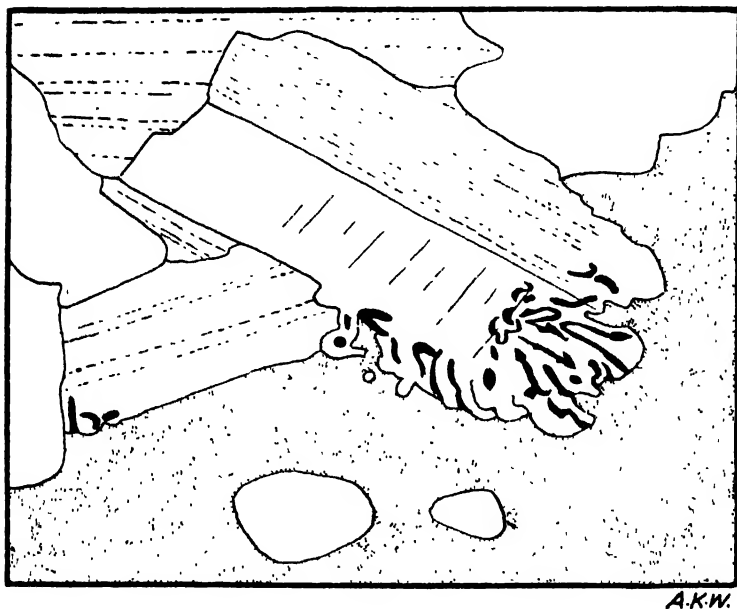


FIG. 73

Myrmekite intergrowths of plagioclase (with twinning indicated) and vermicular quartz (black), penetrating microcline (stippled). The latter contains two rounded blebs of clear quartz. Dancing Cairn, Aberdeen.

which is displaced by the incoming sodium and calcium ions. The potassium probably contributes to the formation of mica as a by-product of myrmekite formation.

The intergrowths are most abundant along plagioclase-orthoclase, or plagioclase-microcline boundaries, so that J. J. Sederholm² has suggested that the juxtaposition of these two contrasted feldspars is essential for the formation of myrmekite. He calls such reaction products formed between two neighbouring solid phases **synantetic**.

¹ "Über Myrmekit," *Tcherm. Min. Petr. Mitt.*, 27 (1908), p. 377.

² "On Synantetic Minerals and Related Phenomena," *Bull. Comm. Géol. Finlande*, No. 48 (1916), p. 134.

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Almost certainly the necessary reagents migrate along intercrystal boundaries, and it is possible that they may not all be entirely deuteritic, for myrmekite is often found in granite adjacent to later basic intrusions, and thus it may be partly of metasomatic origin.

Myrmekite is just one of several phenomena which arise by interaction between adjacent mineral grains. Corona structures, which have a similar origin and are most commonly developed in the gabbroic rocks, are described in due course.

PART III

PETROLOGY OF THE IGNEOUS ROCKS, THEIR SIGNIFICANCE, DISTRIBUTION AND ORIGINS

PRINCIPLES OF ROCK-CLASSIFICATION
AND NOMENCLATURE(1) *General considerations*

THE main purpose in attempting to devise a scheme of rock-classification is to ensure uniformity in nomenclature: it is essential that all petrologists should call the same rock by the same name. Unfortunately this is far from being the case at present, and a considerable part of this book is taken up by explaining (and correcting where necessary) unsystematic and often completely illogical usages. Unfortunately many of the names in common use are of considerable antiquity. At the time they were first introduced only one principle was in vogue: the rock-type took the name of the place where it was discovered and the uniform termination "ite" was added to the place-name, which in some cases was that of a province, in others of a town, village, hamlet or even of a farm, quite regardless of the status either of the place or of the rock itself. As many of these type-localities occur in remote parts of the world where English is not the official language, petrographic nomenclature is heavily burdened with rock-names which look (and sound) outlandish to us.

Names are, of course, essential for these, as for other, natural objects; but we regard it as a duty to keep the number of defined types to a minimum, consistent with clarity. One other important matter arises in this connection. Brevity is not the only consideration: clarity is more important. The important thing to do is to convey to the reader the right idea of the essential features of the rock concerned. As to what are the essential features of a given rock, there is little fear of contradiction in stating that they are two-fold: the lithological character of the rock, depending upon its mineral composition and its texture—the manner of arrangement of the component mineral grains of which the rock is composed; and secondly, its chemical composition. With regard to the relative importance of these two aspects of classification, it depends upon circumstances: for some purposes the facts of mineral composition and texture may be more significant than the chemical data; but in other circumstances the converse is true. The question

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of practicability arises here. Rock analysis is carried out by skilled specialists—it is right outside the scope of the student. On the other hand, with the aid of thin rock-sections it is possible to gain detailed knowledge of the mineral contents of most rocks, while the texture is, of course, observable and in due course interpretable.

(2) *Classification on a chemical basis*

A rock analysis is not in itself very helpful to the student; but it may be made more informative when recast in terms of "standard minerals," the proportions of which provide a basis for comparison with other rocks. The composition so calculated is termed the **norm** of the rock, as distinct from its actual mineral composition, which is spoken of as the **mode**. These two words are often used adjectivally: thus one may speak of "normative quartz" as distinct from "modal quartz." The calculation of the norm is the essential first step in a quantitative scheme of classification of igneous rocks devised by four American petrologists (Cross, Iddings, Pirrson and Washington)¹ and known as the C.I.P.W. classification.

In much the same way Paul Niggli devised a system of calculation expressed as the "Niggli number," which is a symbol based on certain important relationships between selected components. Comparison of the Niggli numbers for two analysed rocks immediately shows the essential similarities and differences between them.

One other matter arises in connection with the calculation of the norm. The results may be expressed by a symbol, and although the latter cannot remove the need for rock-names, it is in effect a concise statement of composition, and as such is far superior to most rock-names, particularly the type of name which tells one nothing of the chemical or mineralogical composition of the rock, but only suggests the name of the place where the original specimens were obtained. Rock-nomenclature is overburdened with "jacupirangites" and "mariupolites." On the other hand it is not particularly inspiring to refer to a rock as $XsM\alpha(25)$, and it would obviously be impossible to use such symbols on a geological map prepared for general use; but to anyone familiar with the excellent scheme devised by S. J. Shand,² the symbol calls up a most comprehensive picture of the essential character of the rock as regards degree of crystallization, the role of silica in it, its coloured minerals, the kinds of feldspar present and the ratio of light to dark minerals. Probably the greatest single advantage of a chemical classification over a mineralogical one is that all rocks of the same composition are automatically put in the

¹ *The Quantitative Classification of Igneous Rocks*, Chicago, 1903.

² *Eruptive Rocks*, London.

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same category, even when the rock is glassy or too finely crystalline for the minerals to be determined.

(3) *Variation in mineral composition*

Rocks which are chemically alike may vary widely in mineral composition on account of differences in conditions during crystallization and cooling. Some of these have been anticipated in the mineralogy section of the book. Hornblende is a case in point.

As shown above, there is a close chemical relationship between amphiboles and pyroxenes, while the physical conditions that favour the crystallization of the one often inhibit the formation of the other. More specifically, if the components of hornblende are available, this mineral will appear in the rock only if the environment is favourable; otherwise almost exactly the same components may appear as augite and a Mg-rich orthopyroxene in association. Similarly, if the chemical compositions are carefully compared it will be seen that the following cases of **heteromorphism** may be expected:—

Biotite and quartz may be represented by orthoclase, hypersthene (or olivine) and magnetite.

Hornblende \rightleftharpoons augite and orthopyroxene.

Hornblende \rightleftharpoons orthopyroxene, olivine, plagioclase.

Augite \rightleftharpoons calcic plagioclase, olivine, diopside, magnetite.

Biotite, orthoclase, hornblende \rightleftharpoons leucite, olivine, anorthite and magnetite.

Biotite, plagioclase \rightleftharpoons orthoclase, nepheline, augite, magnetite.

Plagioclase (sodic), olivine, diopside \rightleftharpoons nepheline, hornblende, plagioclase (calcic).

Orthoclase, nepheline, augite \rightleftharpoons leucite, plagioclase, melilite and olivine.

As we have already seen, the essential minerals in any rock fall into two groups, the felsic and the mafic. The former, comprising the feldspars and the chemically allied feldspathoids together with quartz, are given preference over the mafic constituents comprising the olivines, pyroxenes, amphiboles and dark micas. There are very few rocks which contain no feldspar; in many types feldspars are dominant, and primarily for this reason these minerals play a major part in rock classification and nomenclature.

The **feldspars** are used in the first place to define three major series based on the broad type of feldspar present. For the purposes of classification alkali-feldspar (as defined in the Mineralogy Section) is contrasted with plagioclase (excluding albite). Some kinds of igneous rock contain the former only, in others only plagioclase occurs; but in many rocks both kinds of feldspar occur, in varying

proportions. Opinion is divided as to whether two or three series shall be so defined. In our opinion the advantages of recognizing three major series rather than two outweigh the disadvantages. Broadly the three series are defined as under: (1) by dominance of alkali-feldspars; (2) alkali-feldspar and plagioclase of approximately the same status; (3) plagioclase dominant.

Subdivision may be made, if necessary, on the basis of the variety of alkali-feldspar or of plagioclase occurring in the rocks in the series defined above. Thus the alkali-feldspar may be (a) potassic, (b) sodi-potassic or (c) sodic; while an important distinction is made between rocks containing oligoclase or andesine (An_{-50}) on the one hand, and labradorite, bytownite or (rarely) anorthite (An_{-50}) on the other. Thus the important coarse-grained rocks, diorites and gabbros, and the lava types andesites and basalts are distinguished primarily on the basis of the variety of plagioclase which they contain.

The feldspar-feldspathoid relationship involves recognition of the **silica-saturation** principle (in so far as it affects the felsic minerals).¹ According to this principle all igneous rocks fall into three categories: (1) over-saturated, distinguished by the presence of free (uncombined) silica occurring as quartz; (2) saturated—all the silica is combined and enough is present in the rock to exclude feldspathoids; (3) under-saturated, comprising rocks containing feldspathoids.

The degree of under-saturation is indicated by the feldspar/feldspathoid ratio. Again it is necessary to recognize three categories: (1) with feldspar dominant; (2) with feldspathoid dominant; (3) with feldspathoid exclusive. Further subdivision may be made according to whether the feldspathoid is potassic—leucite or kalsilite—or sodic—nepheline, sodalite, etc.

The role of mafic constituents in rock classification and nomenclature is broadly two-fold: the amount is more important than the kind or kinds present, in most cases. The definitions of many of the common rock-types leave a good deal of latitude as regards both the kind(s) and relative amounts of mafic minerals. Thus andesite contains plagioclase (An_{-50}) in association with a wide variety of mafic minerals which may include enstatite, hypersthene, augite, hornblende or mica. In a few cases a specified mafic mineral *must* be present, for example troctolite consists of olivine and plagioclase (An_{-50}).

The Colour Index is the percentage of mafic minerals in a rock; but more important than the actual figure is the *relative* darkness or lightness of the rock concerned. In this connection it is useful

¹ Shand, S. J., "The Principle of Saturation in Petrology," *Geol. Mag.* (1915), p. 340. See also *Eruptive Rocks*, by the same author.

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to distinguish varieties of a specified type which are (a) above average, (b) average and (c) below average in this respect. Rocks containing more than average mafic minerals are termed **melanocratic**; less than average are **leucocratic**. The corresponding term for the average is seldom used, but is **mesocratic**. It is important to note that these terms are relative only: what is average for one rock-type is certainly not for another, and therefore it is impracticable to choose numerical limits which can apply equally to all the rock series concerned.

By using the contractions leuco- and mela- as prefixes to a particular rock-name greater precision is gained and one rock-name serves instead of three. This course is practicable for those rock-types which consist of a felsic mineral, say nepheline, associated with a mafic constituent, say pyroxene in varying proportions. The ratio of the former to the latter varies actually from essentially "pure" nepheline-rock to "pure" pyroxene rock. The rock-type composed of both components in approximately equal amounts is ijolite. Most known ijolites are notably variable rocks: some specimens are pyroxene-rich, others are nepheline-rich. There is a good deal to be said for distinguishing the former as mela-ijolites and the latter as leuco-ijolites rather than using separate specific names and hence obscuring the gradational relationship which exists between them. The reader should compare the petrographic descriptions of ijolites with those of diorites and gabbros to see how the principle is applied in different cases.

The one-mineral rocks, termed **monomineralic**, should surely be easiest to name and classify; but unhappily this is far from being the case. A bad precedent was established with the well-known and widely distributed rock, which apart from accessory amounts of chromite, consists of olivine only. Logically this should have been named olivinite; but the uninformative term "dunite" was used instead, and is still preferred in many circles. If necessary, more specific varietal names indicating the composition of the olivine may be used instead of the generic term olivinite. For example, certain valuable platinum-bearing olivinites from the Transvaal are actually hortonolites.

Similarly, as noted incidentally above, rocks occur which consist of pyroxenes (with minor accessories) and are termed generically pyroxenites, and specifically hypersthene, bronzite, diallagite and enstatite according to the variety which forms the rock.

By strict analogy monomineralic rocks consisting essentially of amphibole only should be termed amphibolites; but this is one of the inconsistencies which render rock classification a byword with other scientists. Amphibolite as used today is not a monomineralic rock; it does not consist (as the name implies) of amphibole only;

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it is not even an igneous rock. This means, of course, that there is no *collective* name available for these rocks, though the *specific* term hornblendite is.

The position is even worse when the comparable felsic rocks are considered. The most widely distributed type of this kind consists of plagioclase as the sole essential mineral. The term "anorthosite" has been preferred to straightforward "plagioclaseite." Examples of anorthosite consisting of andesine, labradorite or bytownite have been described, and these mineral names could with advantage be used when applicable, with the uniform termination "-ite" added. Albitite is what its name implies; but although it must be regarded as a variety of plagioclaseite it is not usually classed as anorthosite. Again the reader will have no doubts as to what sanidine, nepheline and leucite *ought* to be, but unfortunately none of these names has been applied to rocks of appropriate composition as will be explained in due course.

One other general matter concerning rock-nomenclature must be explained. Many rock-names are compound, consisting of the name of a mineral and that of a rock joined by a hyphen. In many instances the mineral named is an *addition* to the assemblage implied by the rock-name. Thus quartz-syenite contains the typical syenitic mineral assemblage (alkali-feldspar and mafic minerals), with quartz as an additional (minor) component. Similarly olivine-gabbro contains olivine additional to the gabbroic assemblage (plagioclase An₅₀ and pyroxene). However, there are similar compound rock-names in which the named mineral fills a different role. Thus in andesites the plagioclase may be accompanied by mafic minerals in variety: if in a particular specimen the mafic constituent is, say, hypersthene, the rock is a hypersthene-andesite. Unfortunately there is yet a third role which the named mineral may fill. The term "melilite-basalt" (von Rath, 1866) was introduced many years ago for a dark coloured, compact rock superficially resembling basalt; but as soon as the first slide of this rock was examined it should have been seen that the name basalt was inappropriate, for one of the two essential minerals in *all* basalts is absent: there is no plagioclase; its place is taken by melilite. Therefore the role of "melilite" in this compound rock-name is to indicate that it is taking the place of one of the essential minerals in the basaltic assemblage. This is tantamount to classifying a rock on what it does *not* contain—a truly Gilbertian situation that cannot be tolerated.

(4) *Textural considerations and mode of occurrence as factors in classification*

Textural variation in igneous rocks involves several factors, but one is of paramount importance. Variation in degree of crystallinity,

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as evidenced by grain-size, is, after variation in mineral composition, the most important feature contributing to the distinctiveness of rocks. Three grain-size categories are distinguished by the obvious terms "coarse," "medium" and "fine." For general purposes a rock falls in the fine-grained category if the individual components are too small to be seen with the unaided eye. This does not apply, of course, to any large, first generation crystals ("phenocrysts") which may be present, but only to the matrix ("groundmass") in which they may be embedded. Similarly if the groundmass (or the whole rock in the absence of phenocrysts) consists of grains large enough to be seen, but too small for identification, it probably belongs to the medium grain-size category. These are admittedly only rough-and-ready tests. More precise classification involves measurement of grain-size, preferably on thin sections, using arbitrarily chosen numerical limits. The limits in use were chosen by a British Association Committee on Petrographic Nomenclature to correspond as closely as possible with those exhibited in nature by rocks occurring as (1) lava flows, (2) dykes and sills and (3) larger rock bodies of so-called plutonic type. Rock bodies of these three different modes of occurrence constitute the three *phases* of an episode of volcanic activity and represent cooling under three different environmental conditions. Cooling conditions control grain-size, and finely crystalline rocks result from cooling under extrusive or near-surface conditions. It must be remembered that we are classifying rocks—not rock bodies—and rocks, for example basalts, identical in all respects down to the smallest detail can be collected from minor intrusions as well as from lava flows. It should surely be unnecessary to urge that identical rocks must bear the same name regardless of mode of occurrence; but there are still geologists who regard mode of occurrence (not grain size) as being of greater importance than mineral composition and texture in rock classification. Of the many reasons why this mistaken idea cannot be maintained perhaps the most convincing is that it is quite impossible to define "plutonic" with any pretence of scientific precision. Several of the known rock-types have been described, classified and named on material occurring only as ejected blocks, in a few cases the mode of occurrence is unknown or problematical; but this does not prevent those who pay lip-service to this principle from classifying and naming such rocks. Actually all one can go upon is grain-size.

Apart from variation in grain-size (which is of prime importance) textural variations are of minor significance only: they should not be made the excuse for introducing new rock-names—there are more than enough already. A given mineral assemblage in each of the grain-size groups requires a distinctive name; but textural varieties are adequately covered by using appropriate qualifiers.

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Thus nepheline and sanidine together with mafic minerals and accessories in the fine-grained group define the type phonolite. The rock does not have to be porphyritic (though usually it is); but if and when the same mineral assemblage in the same grain-size category is discovered in the form of an equigranular mosaic it is still phonolite, but could be distinguished as "aphyric phonolite." This principle is accepted without question in the case of basalts, and should apply equally to other rock-types defined in terms of their mineral composition and grain size.

(5) *The role of silica in rock classification*

The dominant chemical constituent of igneous rocks is silica (SiO_2), and the silica content has been widely used as an important factor in rock classification. The most siliceous igneous rocks have come to be known as the Acid rocks, a term which has survived from the days when the silicate minerals were regarded as "salts" of various silicic acids. Such a precise and wholly erroneous meaning is no longer attached to the term "Acid," which is used in a general sense to cover all those igneous rocks which contain quartz as an *essential* constituent. At the other extreme of composition are the Basic and Ultrabasic rocks which are, by comparison, silica-poor. These rocks consist largely of minerals containing relatively little silica (such as basic plagioclase and olivine). The latter in particular is characteristic of the Basic rocks in much the same way as quartz is of the Acid ones.

Granitic rocks are dominant in the Acid category and basalts in the Basic category; while between these extremes is a wide range of rock-types including syenites and diorites which have come to be known as Intermediate. As might be expected, this is a rather ill-defined category. In several early editions of this book the igneous rocks were divided into four categories, Acid, Intermediate, Basic and Ultrabasic, separated by arbitrary, but rigidly applied, limits of silica percentage. It was eventually realized, however, that such subdivision, based on a single component, albeit the dominant one, is impracticable and cannot be reconciled with the mineralogical composition of the rocks.¹ Thus numerical values of silica percentage were abandoned in the last edition. Now we have gone farther and abandoned the four-fold classification, though the terms Acid, etc., are used in the text in a qualitative sense, as useful adjectives.

For descriptive purposes the rock-types are grouped much as previously, but without the unnatural limitations imposed by rigid

¹ Wells, A. K., "Silica percentage . . . in igneous-rock classification," *Geol. Mag.*, **28** (1949), p. 688.

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use of silica percentage as a prime factor in classification. Thus it is possible to introduce more flexibility into the grouping of rock-types. Rocks which are closely related genetically (and in the field) need no longer be described in widely separate parts of the book because at some point in the series an arbitrarily chosen silica percentage has been passed. Such divisions are artificial, and complicate rather than simplify classification and nomenclature. Obviously the mineral assemblages occurring in the rocks provide a much more satisfactory basis on which to erect a scheme of classification.

Summarily, the authors' aim throughout is to provide a scheme of classification and nomenclature involving precise definitions based on the facts of mineral composition and grain-size (the most significant textural feature). These are the only features which give distinctiveness and individuality to different rocks. It proves impossible to live up to these ideals without revising the definitions of some rock-types, a number of which are, unfortunately, in common use.

One other important point must be explained: it is impossible to define criteria which are suitable for application to all kinds of igneous rocks. The subdivisions which it seems logical to recognize when classifying the granitic rocks are inapplicable to those at the other end of the compositional scale. We may anticipate criticism by stating our belief that it is necessary to erect subdivisions in certain rock-groups which cannot apply to some others: for example, it is absolutely essential to make fundamental divisions based on silica-saturation when classifying the syenitic rocks and their associates; but obviously this principle is inapplicable to rocks containing quartz as an essential constituent, *i.e.* to granites and the medium and fine-grained equivalent types.

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THE ACID IGNEOUS ROCKS

THE term "Acid" used in the petrographic sense is still of value; but whereas the total silica content of a rock—the measure of its acidity—cannot be seen on inspection (though it is determinable by chemical analysis), the amount of uncombined (free) silica is easily measured in coarse-grained rocks. Thus a given rock may be classified according to the status of the quartz it contains: the latter may be "essential" or "accessory." There is some difference of opinion concerning where to draw the line between the two. In the previous edition the figure favoured was 10 per cent; but there are reasons for believing that this is rather low. Thus the well-known Dresden syenite, which is widely regarded as the type of its kind, contains on average 12 per cent of quartz. If the limit of 10 per cent is rigidly enforced, this typical syenite must be classified as granite—a position which is obviously anomalous. F. Chayes, who has made detailed studies of the mineral composition of granites and who therefore possesses special knowledge of this subject, advocates applying the term granite only to coarse-grained igneous rocks containing between 20 and 40 per cent quartz. The average for true (alkali-) granites is 30 per cent and these figures give an even spread around the average; but we do not propose using a fixed upper limit—there is no category of igneous rocks more highly silicated than granites and therefore an upper limit to the quartz-content of these rocks is not necessary.

As already explained, three categories of quartz-rich rocks are defined in terms of their feldspar content, and each of these comprises rocks of coarse, medium and fine grain. There are thus three suites of three types each, and there are two ways in which they could be grouped for description: the alkali granites together with their medium and fine-grained equivalents could be dealt with first, followed by the adamellite suite and finally by the granodiorite suite. Alternatively the three coarse-grained types could be described first, then the three medium-grained types, followed finally by the fine-grained types. There are advantages and disadvantages

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both ways—a certain amount of repetition is inevitable. The latter course is used in the account which follows.

The general relationships between the several types of rocks containing essential quartz are shown in the appended table. The coarse-grained rocks in all three categories are collectively, and admittedly rather loosely, termed “granites”; and for ordinary purposes this practice may be condoned; but for more accurate petrological purposes it seems best to restrict the term granite to those types which consist essentially of quartz, *alkali*-feldspars and minor quantities of mafic minerals.

Grain Size	Alkali-Feldspar > $\frac{1}{2}$	Alkali-Feldspar and Plagioclase each between $\frac{1}{2}$ and $\frac{1}{3}$	Plagioclase > $\frac{1}{2}$
Coarse	ALKALI-GRANITES	ADAMELLITES	GRANODIORITES
Medium	ALKALI-MICROGRANITES	MICRO-ADAMELLITES	MICRO-GRANODIORITES
Fine	ALKALI-RHYOLITES	RHYO-DACITES OR TOSCANITES	DACITES

In former editions potassic granites (“potash granites”) were separated from sodic granites on the basis of the dominance of K'-feldspar in the former and Na'-feldspar in the latter. This is actually impracticable (except on the basis of normative feldspar), as the two feldspars are usually inextricably involved as perthitic intergrowths. The majority of these rocks are sodipotassic; but in the most sodic varieties not only is the feldspar chiefly albite or albite-rich antiperthite, but the dark minerals are distinctive NaFe-rich pyroxenes and/or amphiboles. There is full justification, therefore, for considering the sodic granites separately.

(1) ALKALI GRANITES

Potassic and Sodi-potassic Granites

Arising out of their experimental studies of the system orthoclase, albite, quartz and water, Tuttle and Bowen have described the relationships between the feldspars in alkali-granites, and have suggested that significant differences in cooling history are indicated by these relationships. Certain granites contain only one kind of feldspar, an orthoclase-albite intergrowth. These one-feldspar

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granites¹ are probably restricted to high level, sub-volcanic environments, where cooling would be relatively rapid, and therefore the degree of unmixing of the two components in the perthitic intergrowths would be slight. The component minerals exhibit high-temperature optics. Certain British Tertiary granites in the Western Isles of Scotland are of this type and are accepted as being magmatic.

By contrast, other alkali-granites contain two or more kinds of feldspar: albite (not intergrown), orthoclase or (more commonly) microcline may accompany one of the Or-Ab intergrowths. The significance of these differences is discussed more fully below, when the stability relationships are explained. Meantime it will suffice to point out that, provided the feldspars were initially high-temperature, homogeneous types, it is obvious that unmixing has gone much farther in two-feldspar, than in one-feldspar granites, presumably because of the less steep temperature-gradient with a correspondingly long cooling period.

Quartz averages perhaps 30 per cent of the whole, and varies in its relationship to the feldspars. It is invariably the low-temperature (α) form and builds irregular shaped, composite grains, sometimes interstitial to the feldspar sometimes lobed into it in a manner suggesting replacement (Fig. 74). In the more potassic varieties containing large microclines, some quartz forms relatively small rounded inclusions in the latter; but some is moulded upon the microcline. Finally, quartz frequently participates in the formation of the delicate intergrowth occurring along plagioclase-microcline boundaries, and known as myrmekite (Fig. 73).

Inclusions in the quartz are ubiquitous, and in thin sections are seen to consist of trains of minute bubbles sealed-in in healed fractures in the crystals. Minute acicular crystals, identified (by analogy with more robust, megascopic specimens) as rutile, may occur in considerable numbers—but they have to be searched for in the thin sections. As these rutiles are restricted to the quartz grains, it may be inferred that their substance was originally held in solid solution in the quartz, but separated from the latter by unmixing at the appropriate temperature. Such rutile needles indicate that the quartz in which they occur must have originated as the high-temperature form and thus provide valuable evidence regarding temperatures of crystallization of the granites concerned.

Apart from rutile a little iron-ore may occur, and may well be the only accessory visible in a given thin section; but by crushing rock-samples and using the "heavy-mineral technique" of concentration and separation, a surprisingly large variety of the rarer accessories may be obtained. These are referred to below; but it

¹ One-feldspar granites are the "*hypersolvus granites*"; and two-feldspar granites the "*subsolvus granites*" of Tuttle and Bowen.

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may be noted that the biotite may be crowded with pleochroic haloes formed by irradiation from such radio-active minerals as zircon and xenotime.

All true alkali-granites are relatively deficient in mafic minerals, some more so than others. These are distinguished as **leucogranites**—they are very light-coloured with scattered grains of dark mica. A facies of the Dartmoor granite from Wittabarrow, Devonshire,

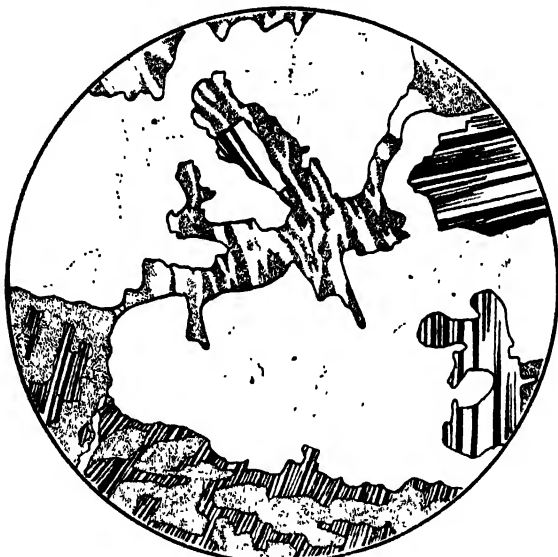


FIG. 74

Alkali-leucogranite, Mount Mado, Jersey, C.I.

Reproduced by permission of the Council of the Geologists' Association, London.

The minerals shown are quartz and two kinds of alkali-feldspar, microperthite and water-clear albite: the rock is a two-feldspar leucogranite. To show the feldspar relationships to advantage the feldspars are drawn as between crossed polarizers, but the quartz is left clear.

consists of alkali-feldspar and quartz making up 97 per cent of the rock, with only 3 per cent of mafic and accessory minerals.

Micas are the only mafic minerals commonly seen in these granites and include biotite and/or muscovite often closely associated. The biotite is often partially altered to chlorite.

Examples of sodi-potassic granites are not uncommon among the Armorican granites of south-western England and the Channel Islands (Fig. 74).

It is certainly no coincidence that the alkali-granites seem to be particularly prone to alteration by late-stage or pneumatolytic alteration-processes, as described below. Incipient alteration of

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these types may cause small flecks of sericite or paragonite to form, often along cleavages, while kaolinization may cause clouding and ultimately opacity of the feldspars.

A distinctive type, of somewhat doubtful affinities, is **charnockite**, described from a locality in Madras. The type-rock contains quartz 40, microcline 48, oligoclase 6, hypersthene 3, biotite 1, and magnetite 2 per cent. In hand specimens the quartz appears bluish, while in thin section it is seen to be charged with extraordinary numbers of minute acicular crystals, identified as rutile, occurring in hundreds of thousands per cubic centimetre. Apart from this, the outstanding feature is the occurrence of distinctly pleochroic hypersthene. This links charnockite with a series of hypersthene-bearing rocks of varying composition—the so-called Charnockite Series,¹ which spans the whole range of composition, from Acid to Ultrabasic, and builds complexes known to occur in many parts of the world: in India (Madras is the type locality), Ceylon; Africa² including Uganda, the former Belgian Congo, Natal, the Central Sahara and Madagascar; also in the Fenno-Scandian Shield (Finland) and the Crystalline Shield area of the Ukraine and Siberia. In America charnockites appear to be more restricted, but do occur in eastern Brazil.

The constant feature is the presence of pleochroic orthopyroxene, of iron-rich type, associated with smaller amounts of augite (rarely), a somewhat distinctive hornblende, or biotite. The most distinctive accessory, not invariably present, is garnet, of pyrope-almandine type.

Among the rocks broadly included in the term charnockite are varieties characterized by a preponderance of microcline—these are the true charnockites, very closely matching the specimens originally described by Holland (1900) from the type locality. In addition, however, others contain two feldspars, a K'-type and a plagioclase in the proportions which would place them among the adamellites of our general classification. They are hypersthene-adamellites, showing the mineralogical and textural distinctive features which link them with the Charnockite Series, and are therefore distinguished as "charnockitic adamellites." Finally, in other specimens plagioclase within the andesine range (An_{35}) is dominant over microcline, and the rocks fall among the granodiorites. Charnockitic granodiorites have been given the name "**enderbite**," after Enderby Land, the type-locality in Antarctica.³

¹ Holland, T. H., "The Charnockite Series . . .," *Mem. Geol. Surv., India*, Vol. 28, Part II (1900), p. 162; and Rao B. Rama, *Bull. Mysore Geol. Dept.*, No. 18 (1945).

² Howie, R. A., "African charnockites and related rocks," *Service Geol. Congo Belge, Bull.*, No. 8, Fasc. 2 (1958).

³ Tilley, C. E., "Enderbite, a new member of the Charnockite Series," *Geol. Mag.*, 73 (1936), pp. 312-16.

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It seems to be generally agreed that these charnockitic complexes were originally magmatic. In at least some occurrences they show an apparently intrusive relationship to the country-rock, which consists of metamorphic types belonging to the granulite facies.

As they are demonstrably not younger than the general metamorphism if, at the onset of granulitizing conditions, they had been already solid *rocks*, their original mineral assemblage and textures would be readjusted to the new conditions, and the rocks would be classified as metamorphic. There is another possibility, however: the magma may have been intruded *during* the metamorphism, so that the *original* consolidation took place in a granulitizing environment.

It is a matter of no practical importance, though of academic interest to prove, in a given instance, which of the two processes applied. If, as appears likely, some charnockites were originally igneous *rocks*, and were reconstructed under the appropriate metamorphic conditions, there is a chance that some parts of the rock-body would not have achieved complete equilibrium, and that relics of the original mineral assemblage and textures would have survived and still be recognizable. This is claimed to be the case with certain charnockitic rocks occurring in the Enisey range, Siberia, and in which vestiges of basic igneous rocks are discernible.

In any case charnockites and enderbites are border-line cases: their petrographic interest is unquestioned; they provide an opportunity for studying an interesting example of heteromorphism. The problem of their mode of origin is of secondary interest.

Sodic Granites

True sodic granites are rare, but those which do occur are very distinctive rocks, particularly as regards their coloured silicates. Typical examples occur in ring-complexes in Nigeria¹ and the Sudan. Some of the rock-types building these complexes are ordinary biotite-granites; but of much greater interest are associated sodic rock-bodies of plutonic, extrusive and hypabyssal types, characterized by aegirine and/or riebeckite. The distinctiveness of the mineral suite is, of course, merely the expression of the chemical composition of the magma-fraction from which it was formed, and which has been shown by analysis to have been relatively rich in sodium but poor in Ca, Al and Mg. Deficiency in the first two affected principally the feldspar association in the rocks: the anorthite content is exceptionally low, the plagioclase being nearly

¹ Jacobson, R. R. E., Macleod, W. N., and Black, R., "Ring-Complexes in the Younger Granite Province of Northern Nigeria," *Mem. No. 1, Geol. Soc. London* (1958), pp. 72, *et seq.*

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ANALYSES OF ALKALI-GRANITES

	I. Charnockite.	II. Sodic leuco- granite, Mountsorrel	III. Riebeckite- granite.	IV. Aegirine- granite, Rockall.	V. Potassic leucogranite, Dartmoor.	VI. Potassic granite, Dartmoor.	VII. Potassi-sodic granite, Dartmoor.	VIII. Potassic granite, Lamorna.
SiO ₂	70.65	76.70	76.25	70.31	73.16	73.66	71.69	74.69
Al ₂ O ₃	15.09	12.58	10.86	7.53	13.95	13.81	14.03	16.21
Fe ₂ O ₃	0.80	0.10	1.23	8.32	0.03	0.21	0.57	tr.
FeO	1.53	2.09	0.76	2.44	0.47	1.51	1.93	1.16
MgO	0.53	0.65	0.18	0.02	tr.	0.45	0.66	0.48
CaO	2.66	1.10	0.37	0.35	0.43	0.67	1.49	0.28
Na ₂ O	2.99	4.90	4.68	5.26	2.57	2.89	3.03	1.18
K ₂ O	4.69	0.52	4.65	4.19	8.16	5.02	4.59	3.64
H ₂ O	0.65	0.85	0.50	0.43	0.64	1.66	1.76	1.23
TiO ₂	0.46	0.20	0.11	0.26	0.04	0.16	0.33	—
Rest	0.10	0.14	0.36	0.54	0.17	0.31	0.29	0.68
	100.15	99.83	99.95	99.65	99.62	100.35	100.37	99.55

- I. Charnockite, Madras, India (Anal. J. H. Scoon), R. A. Howie, *Trans. Ed. Roy. Soc.*, **62** (1955), p. 98.
- II. Sodic leucogranite, marginal facies of Mountsorrel granodiorite, Charnwood Forest, Leicestershire (Anal. W. H. Herdsman), *Geol. Mag.* (1934), p. 1.
- III. Riebeckite-aegirine granite, Nigeria (Anal. R. O. Roberts), *Geol. Soc. Lond. Mem.*, **i** (1958), p. 17.
- IV. Aegirine-granite, Rockall, P. A. Sabine, *Bull. Geol. Surv. G.B.*, **16** (1960), pp. 156-78.
- V. Potassic leucogranite ('aplogranite'), Widdabarrow, Dartmoor (Anal. H. F. Harwood), *Min. Mag.*, **20** (1923), p. 41.
- VI. Potassic granite, Haytor East Quarry, Dartmoor (Anal. H. F. Harwood, *op. cit.*).
- VII. Potassi-sodic granite, Saddle Tor, Dartmoor (Anal. H. F. Harwood, *op. cit.*).
- VIII. Potassic granite, Lamorna, Cornwall (Anal. W. Pollard).

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pure albite. Further, among coloured silicates the more normal biotite and hornblende were ruled out, their places being occupied by non-aluminous and calcium-free aegirine and riebeckite, sometimes alone, but often in the closest reaction relationship, the latter being usually external to the former. One other special feature of these rocks may be noted: they are rich in an assortment of the rarer accessory minerals, including some of considerable economic importance. A high content of fluorine finds expression in the presence not only of fluorite, but also of cryolite, the corresponding fluoride of sodium, occurring in exploitable amounts in one known riebeckite-granite. Thorite, monazite and xenotime occur, as well as radio-active pyrochlore, visible even in hand-specimens as small honey-coloured octahedra, and valuable as a source of the element Niobium. Astrophyllite, seen in thin sections as bright yellow, micaceous-looking aggregates, locally becomes so important in the marginal facies of some riebeckite-granites as to rank as an essential constituent. Nearer to hand "peralkaline" granites build the islet of Rockall¹ in the North Atlantic, the dominant type being one carrying both aegirine and riebeckite. A melanocratic facies has been called **rockallite**, in which the colour-index is very high for granitic rocks—39; the feldspar is albite, 23, while quartz totals 38 per cent.

As defined above, sodic granites are not necessarily excessively alkaline: with rather more Al and less Fe, a type occurs which is no less strongly sodic, but is more ordinary in its coloured mineral content. In such rocks the whole of the Na is locked up in the feldspar, which, by definition, may be albite, but is commonly either micropertite, antiperthite, or cryptoperthite. These are the dominant components, associated, of course, with the requisite amount of quartz, with biotite and the usual accessories.

(2) ADAMELLITES

In this category are included those granitic rocks in which Ca-bearing plagioclase accompanies a potassic feldspar in approximately equal amounts: neither the potassic feldspar nor the plagioclase should exceed two-thirds of the total feldspar present. Thus, compared with the alkali-granites, adamellites are distinctive through the increasing importance of Ca-ions in the feldspar. Generally the plagioclase lies within the oligoclase range, but may be andesine. The essential quartz, alkali-feldspar and plagioclase are accompanied by biotite in some adamellites, but by biotite and common hornblende in others, while, as noted above (p. 192), some of the rocks included in the Charnockite Series are charnockitic adamellites.

¹ Sabine, P. A., "The Geology of Rockall, North Atlantic," *Bull. Geol. Surv. G.B.*, 16 (1960), pp. 156-78.

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ANALYSES OF ADAMELLITES AND GRANODIORITES

	I. Adamellite, Abriachan.	II. Adamellite, Rubislaw.	III. Adamellite, Shap Fell.	IV. Adamellite, Aberlour.	V. Trondhjemite Average.	VI. Granodiorite, Glen Barvie.	VII. Granodiorite, Moor of Rannoch.	VIII. Granodiorite, Nevada City.
SiO ₂	71.25	69.01	68.88	68.01	70.27	67.86	66.91	66.65
Al ₂ O ₃	18.03	17.74	14.17	17.34	16.01	16.62	15.09	16.15
Fe ₂ O ₃	1.29	0.97	0.87	3.66	0.89	0.81	1.70	1.52
FeO	0.34	2.05	1.66	—	1.33	2.16	2.13	2.36
MgO	0.38	0.48	1.33	0.77	0.77	1.52	2.02	1.74
CaO	2.61	1.95	2.32	2.06	2.98	2.13	3.32	4.53
Na ₂ O	2.25	2.73	3.30	2.05	5.31	5.08	4.16	3.40
K ₂ O	3.09	3.94	5.06	4.41	1.53	2.54	3.16	2.65
H ₂ O	0.82	1.18	1.25	0.64	0.55	0.95	0.70	0.90
TiO ₂	—	—	0.62	0.83	0.22	0.46	0.67	0.38
Other constituents	0.13	—	0.51	0.48	0.14	0.06	0.17	0.29
	100.19	100.05	99.97	100.25	100.00	100.19	100.03	100.57

- I. Adamellite, Abriachan, Loch Ness, Scotland (W. Mackie).
 II. Adamellite, Rubislaw, Scotland (W. Mackie).
 III. Adamellite (Stage II Granite), Shap Fell, Cumberland (Anal. H. F. Harwood), *Proc. Geol. Assoc.* (1928), p. 311.
 IV. Adamellite, Ruthre Granite, Aberlour, near Ben Rinnes, Strathspey.
 V. Trondhjemite, average of five analyses quoted by Johansen (1933), p. 385.
 VI. Granodiorite (sodic type), Glen Barvie, Scotland (Anal. W. H. Herdman), *N. Holgate, Quar. Journ. Geol. Soc.*, 106 (1951), p. 438.
 VII. Granodiorite, Moor of Rannoch, Scotland (Anal. E. G. Radley).
 VIII. Granodiorite, Nevada City, California (W. Lindgren).

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They contain essential quartz, K'-feldspar and plagioclase in approximately equal quantities, and pleochroic hypersthene. In the definition of adamellite no limitation is placed upon the content of dark minerals; but as a rule the colour-index is rather higher than in alkali-granite, though lower than in granodiorites.

The name **adamellite** was originally suggested for a type now distinguished as tonalite, but was redefined by Brögger (1895) in substantially the sense in which we use it here. The type locality is the Adamello Complex in the Tyrol; but adamellite is a widely distributed type, though often referred to merely as biotite-granite or biotite-hornblende-granite. A well-known British example occurs at Shap Fell¹ in Westmorland. Part of the complex consists of a particularly handsome rock characterized by numbers of large pinkish phenocrysts of orthoclase, embedded in a granular aggregate of white oligoclase, quartz and biotite. Micrometric analysis² shows this adamellite to contain approximately quartz 24, orthoclase 36, oligoclase 34 and biotite 6 per cent.

It may be noted that the "Intermediate" equivalents of adamellites are monzonites, which differ chiefly in being quartz-free, or contain only subordinate amounts of this mineral. Some petrologists therefore use the term "quartz-monzonite" instead of adamellite.

③ GRANODIORITES

Of all the coarse-grained, quartz-rich rocks, granodiorites are quantitatively the most important: indeed they are far more widespread than all the coarse-grained members of the Intermediate and Basic Clans combined. In existing records, however, care must be taken to see the exact sense in which the name is being used—usage is varied, and there is a good deal of confusion as between granodiorite, tonalite and quartz-mica-diorite. In all of these types plagioclase is dominant, to the extent of at least two-thirds of the total feldspar. While granodiorite is quartz-rich, however, tonalite and quartz-mica-diorite contain quartz as an **accessory** only, not as an **essential** component.

Granodiorites, then, are coarse-grained igneous rocks containing essential quartz with plagioclase dominant, though alkali-feldspar may occur, but must not exceed one-third of the total feldspar content. These felsic minerals are accompanied by a varying proportion of coloured silicates and accessories, of which biotite and hornblende are almost constantly present in the former, and sphene, apatite and magnetite in the latter category.

This is almost exactly the original sense in which the name was

¹ Grantham, D. R., "Petrology of the Shap Granite," *Proc. Geol. Assoc.*, 39 (1928), p. 299.

² Holmes, A., *Petrographic Methods and Calculations* (1921), pp. 594-9.

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first used (Becker, 1892, in conjunction with H. W. Turner¹ and W. Lindgren) as applied to certain rocks of the Sierra Nevada range. Thus defined, granodiorite transgresses the silica-percentage boundary (at 66 per cent) formerly used to separate the Acid from the Intermediate igneous rocks: there are as many granodiorites with less than 66 per cent as there are with more than this amount of total silica. The average is 66 to 67 per cent, and of this about 22 per cent is quartz. There is a perfect gradation into tonalite, which differs only in the subordination of quartz.

Granodiorites in which alkali-feldspar is completely (or almost completely) suppressed, have been termed **trondhjemites** (Goldschmidt, 1916), and consist of plagioclase of the appropriate range of composition for granodioritic rocks—oligoclase to andesine—together with quartz and small quantities of biotite, sometimes proxied by hornblende or pyroxene. With an average SiO_2 percentage of over 70, and 20 to 30 per cent of quartz, it is appropriate to include them as orthoclase-free granodiorites, which in mineral composition are very close to quartz-rich tonalites.

It is in the enormous "granitic" batholithic complexes of the mountain ranges of western North America that granodiorite, together with the closely similar tonalite, really "holds the stage." The two together have been proved by detailed mapping to occupy the greater part of the surface area of the southern California batholith—3,500 out of 4,000 square miles mapped.²

In Britain granodiorites are far less spectacular, of course, but nevertheless are well represented, with the main Donegal granodiorite at the head of the list, occupying an area of 30 by 5 miles.³

Many loosely named "hornblende-biotite granites" are granodiorites. As an example we may quote the Mountsorrel intrusion from Charnwood Forest, Leicestershire. The major part of this intrusion is granodiorite containing, on average, quartz 22·6, alkali-feldspar 19·7, plagioclase 46·8, biotite 5·8, hornblende 2·9, and magnetite 2·2 per cent. Similar types occur also among the Caledonian and Devonian granitic complexes of the Highlands and Southern Uplands of Scotland, as in the Moor of Rannoch and Ben Cruachan complexes. The average composition of granodiorites from the Garabal Hill–Glen Fyne complex⁴ is all but identical with the Mountsorrel rock quoted above, but there are striking textural

¹ "The Rocks of the Sierra Nevada," *14th Ann. Rep. U.S. Geol. Surv.* (1894), pp. 478 and 482.

² Larsen, E. S., "Batholithic and associated rocks of Corona . . . California," *Mem. Geol. Surv. Amer.*, 29 (1948).

³ Pitcher, W. S., Read, H. H., and others, "The Main Donegal Granite," *Quar. Journ. Geol. Soc.*, 114 (1958), pp. 259–305.

⁴ Nockolds, S. R., "The Garabal Hill–Glen Fyne Igneous Complex," *Quar. Journ. Geol. Soc.*, 96 (1940), p. 451.

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differences: the Scottish type includes strongly porphyritic varieties, in which the phenocrysts consist of alkali-feldspar (microcline-microperthite) while plagioclase forms subhedral grains, usually zoned, and associated with quartz, hornblende, mica and prominent pale brown sphene.

THE TEXTURES OF GRANITES, ADAMELLITES AND GRANODIORITES

The term "granitic" used in a textural sense implies no more than that the rock concerned is granular, *i.e.*, like a granite. Typically, none of the major constituents is bounded by crystal faces: their shapes have been determined by mutual interference during growth. This most typical of the wide range of granitic textures is sometimes termed "xenomorphic granular." It sometimes happens that the feldspars tend to exhibit crystal faces, though the quartz is again anhedral. For this variant, the term "hypidiomorphic granular" is used.

The handsomest granites are undoubtedly the **porphyritic** types in which phenocrysts of white, grey or red feldspar occur embedded in a groundmass that may be identical in texture with an ordinary aphyric granite. The phenocrysts may be strongly zoned, and attain in some cases among the West of England granites to 7 inches by 5 inches, measured on the side-pinacoid faces. They frequently lie in parallel orientation and thus provide significant data in structural studies of granitic rock-bodies. Instances of such granites have already been mentioned in the foregoing account. In Britain the Shap adamellite and the porphyritic variety of the Garabal Hill-Glen Fyne granodiorite are good examples; but the most striking specimens are obtained from the "giant granite" from Dartmoor.

The great majority of granites may be referred to one or other of these two main textural types. There remain for consideration certain varieties which call for more detailed consideration. **Orbicular granites**¹ contain "orbs" of various sizes—they may measure several inches in diameter—embedded in a matrix of normal granitic texture. The "orbs" are, in fact, variously shaped, and they exhibit extraordinary rhythmic banding (Fig. 75). Each of the orbs contains a core, which in some instances is of the same composition as the matrix in which they are embedded, but in other cases it is completely different. Again, although the texture of the matrix is sometimes normal granitic, it may show a continuation of the successional crystallization of the component minerals, in the sense that the interspaces between the orbs are filled largely with feldspar, but with

¹ Johanssen, A., *Descriptive Petrography of the Igneous Rocks*, Vol. iii, p. 248. See particularly the photographs of a range of different types of orbicular granites.

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mica (biotite) in the centre, representing the last mineral to crystallize. Most specimens so far described occur in Finland,¹ and some of them are objects of outstanding petrological interest. Much has been written concerning the significance of orbicular structure. Nuclei appear to be essential, and may be either xenoliths of wall-rock or

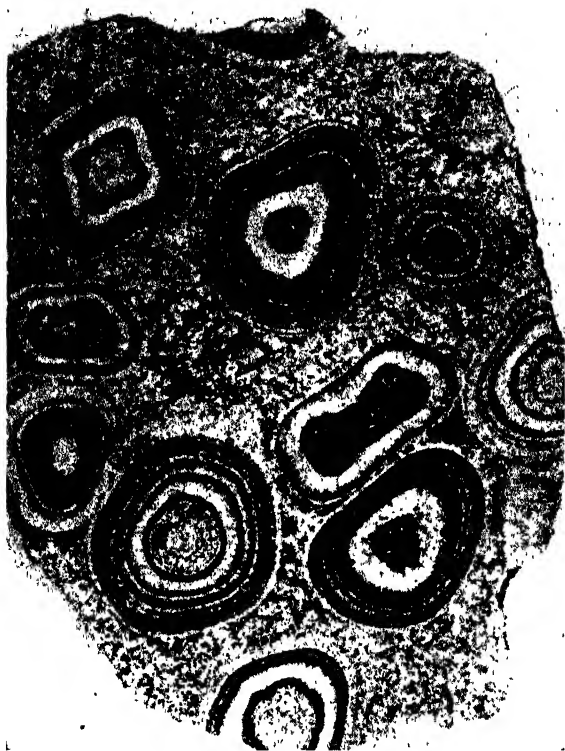


FIG. 75

Orbicular granite, Kangasala, Finland. $\times \frac{1}{2}$

Photo by A. V. Weatherhead.

cognate xenoliths of granite, *i.e.* material of early consolidation broken up by, and incorporated in, later granitic magma. The magma may have been highly viscous, and diffusion consequently slow. This would favour successional rhythmic crystallization around the nuclei.

The rapakivi texture also was originally described from Finnish granites. In typical specimens large flesh-coloured potassic feldspars

¹ See Sederholm, J. J., "On Orbicular Granites . . .," *Bull. Com. Géol. Finlande*, No. 83 (1928).

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form rounded crystals a few centimetres in diameter, and are mantled with white sodic plagioclase, in some cases rhythmically zoned with orthoclase. These feldspars are embedded in a matrix of normal texture, but consisting chiefly of quartz and coloured minerals. Recently the term "rapakivi" has been applied to granites and porphyritic microgranites which contain phenocrysts of reddish orthoclase with narrow mantles of whitish sodic feldspar; for example, some facies of the Shap, Dartmoor and Jersey granites and associated minor intrusions. It is customary in these days to regard phenocrysts with suspicion—the possibility that they originated elsewhere than in the magma represented by the matrix in which they are now embedded must be given due consideration. With the Finnish rapakivi granites this may well have been the case, for the rounded form of the "phenocrysts" suggests, but does not prove, magmatic corrosion. Growing crystals develop plane surfaces only if surface-tension conditions are appropriate, otherwise curved surfaces develop. Therefore a rounded shape may be original. Genuine rapakivi granites are orthoclase-rich, therefore orthoclase crystallizes first, and with slow cooling large phenocrysts may be built up. With falling temperature quartz begins to crystallize, and subsequently all three components—orthoclase, quartz and sodic plagioclase—may crystallize simultaneously. Some of the plagioclase would be deposited round the orthoclase nuclei, but some of it might form independent crystals, particularly if crystallization was accelerated. It has been suggested that continuous build-up of vapour-pressure may have led to the ultimate fracturing of the "roof" above the granite, allowing the water vapour to escape, with the consequence of the setting up of many new centres of crystallization. Much of this cannot be proved, of course, but here is a suggested mode of origin which would account for the distinctive features of these interesting rocks.

The **graphic** or **runic texture** is one of the most distinctive shown by granitic rocks, and is particularly characteristic of the pegmatitic facies. Both terms have reference to the marked resemblance of the small quartz "hieroglyphs" to runic characters, which show up clearly against the background microcline in which they are embedded (Fig. 76). This texture is considered more fully under the heading "pegmatites" below.

Finally, some granites contain small irregular cavities lined with well-terminated crystals of the normal constituents of the rock, accompanied by some of the rarer accessories. Such granites are said to be **miarolitic**. In Britain the granites of the Mourne Mountains in Ireland and Lundy Island off the Somerset coast provide typical specimens. Smoky quartzes, well-terminated feldspars, "books" of white mica as well as rarer apatite and topaz crystals occur in the

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manner described. Again there is a link with pegmatites, for the most likely cause of the phenomenon is the existence in the magma of "pockets" of gas which included the fluorine sealed up in the white mica, topaz and apatite.

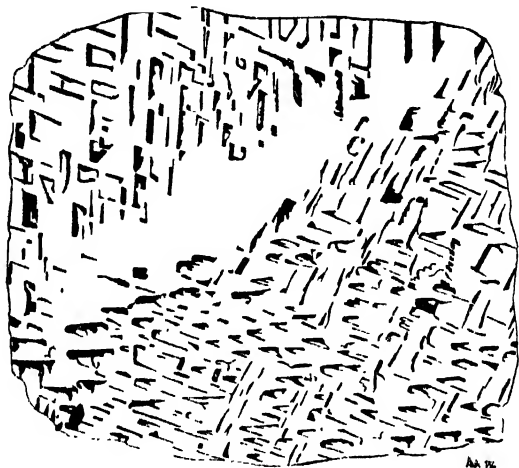


FIG. 76

Sketch of hand-specimen of runite (graphic granite). Microcline plain; quartz hieroglyphs in black, showing shapes controlled by three directions in the host-mineral. *Natural size.*

GRANITE-PEGMATITES AND APLITES

The rocks we have studied so far represent the major portion of the granite magma: the final residuum of large intrusions is naturally rich in the fugitive constituents of the magma, and by the freezing of this residuum rocks are produced which differ markedly in both mineral composition and texture from the normal granite. In texture they fall into two contrasted types: the first of relatively coarse grain; the second relatively fine. The former are the granite-pegmatites, and the latter granite-aplites.

Granite Pegmatites

Few rocks are so well endowed with the power of capturing the imagination as the pegmatites. They are extremely coarse-grained and contain not only the world's largest crystals, but also the choicest mineral specimens. Further, some pegmatites are of considerable economic importance, since they are formed by the crystallization of residual solutions in which there is often a marked concentration of rare elements.

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The place that the pegmatite fluids hold in the cooling history of a granite mass has already been mentioned (p. 170). That the main part of the crystallization is completed before the formation of the pegmatitic residuum is obvious from the relationship between pegmatitic bodies and the main granite masses with which they are associated. Most pegmatites occur in the form of irregular segregations, veins or small dykes or sills, which are especially abundant in the marginal parts of the parent granite, or in the country rock surrounding it, though some pegmatite veins occur at a distance from the main intrusion, and appear to have no visible connection with it.¹

Pegmatite dykes do not, however, form an invariable part of all exposed granite masses: for although most of the latter contain occasional ultra-coarse-grained schlieren or clots, some are associated with a profusion of pegmatitic veins and dykes. The Precambrian granites seem particularly well endowed in this respect; but pegmatites are very poorly represented in association with many granites of later periods, for example the great granitic batholiths of the Andes. In many instances the pegmatite dykes run conformably with the strike of the country rock, and in such abundance as to simulate a dyke-swarm.² Pegmatites are characteristic of regions of compression, not of tension, as is the case with true dykes. Further, unlike true dykes, pegmatites are normally quite short, irregular in form, and rarely parallel-sided. The extent to which pegmatites are formed by *in situ* replacement remains an open question.

Mineral Composition

Alkali-feldspar is the dominant constituent of all granitic pegmatites. In those classified as **simple pegmatites** it is generally microcline-micropertthite, associated with quartz and white mica. Among the abundant and varied accessory minerals are those normally occurring in granite, together with others which are often regarded as "pneumatolytic," such as tourmaline, topaz and other fluorine-bearing minerals and various ores, among which cassiterite is one of the most important.

Less commonly pegmatites consist of a much more varied mineral assemblage. In these **complex pegmatites** albite, or clevelandite, is an important constituent, often exceeding microcline in amount, and is accompanied by a suite of lithium-bearing minerals including red and green parti-coloured tourmalines, spodumene and lithja-micas,

¹ Anderson, Olaf, "Discussion of Certain Phases of the Genesis of Pegmatite," *Norsk. Geol. Tidsskr.*, Vol. 12 (1931), p. 1.

² See, e.g., the account given by Gevers, T. W., and Frommurze, H. T., "The Tin-bearing Pegmatites of the Erongo Area, S.W. Africa," *Trans. Geol. Soc. S. Africa*, 32 (1929), p. 111.

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also beryl and minerals containing niobium, tantalum and other rare elements. The special interest attaching to the complex pegmatites is discussed below, under the heading "paragenesis."

Textural Features

One of the most characteristic features of granite-pegmatites is the development of the graphic or runic texture, resulting from the close intergrowth of feldspar—generally microcline-microperthite and quartz. It was to this structure that Haüy originally applied the term "pegmatite," although in later years this has given place to "graphic" or "runic."¹

Graphic granite presents many striking features and as many problems. In the most regular examples the quartz takes the form of sub-parallel, elongated prisms which pass through the feldspar, and present the characteristic hexagonal cross-sections, often flattened and distorted in the manner of ordinary quartz crystals. Fersmann has shown that the vertical axes of the quartz crystals are so inclined as to make an angle of approximately 70 degrees with the C-axis of the host feldspar. However, other investigators have been unable to confirm this regularity suggested by "Fersmann's Law." Apart from their attitude, it has been found that the proportion of quartz to feldspar is fairly constant, at about 30 to 70 per cent. These various features strongly suggest that the structure results from the crystallization of a eutectic or cotectic mixture of the two components. Despite this, however, origin by replacement of the feldspar by quartz has been urged by certain authors.²

The most striking textural feature of pegmatites is their extraordinary coarseness of grain. Impressive examples are afforded by a beryl crystal 19 feet, and a spodumene 47 feet long, associated with microcline-perthites several feet through, discovered in a pegmatite at Keystone, in the Black Hills, South Dakota. The growth of such large crystals must result primarily from the very low viscosity of the pegmatitic fluid, caused by the presence in it of abundant fluorine, magmatic water and other substances of low atomic and molecular weights. T. Quirke and H. Kremers³ have suggested that the movement of fluid of constantly varying concentration and temperature through the interstices of the crystallizing pegmatite may be an important contributory factor in building up large crystals: a constantly replenished mother liquor could, in this way, largely surround each growing crystal. Occasionally, however, details of

¹ Haüy's original term has survived, however, in "micropegmatite," which is a comparable structure, though on a finer scale, seen in the interstitial quartz-feldspar intergrowth in certain quartz-gabbros and quartz-dolerites.

² Schaller, W., "Mineral Replacements in Pegmatites," *Amer. Min.*, 12 (1927), p. 59; also Wahlstrom, E., "Graphic Granite," *Amer. Min.*, 24 (1939), p. 681.

³ "Pegmatite Crystallization," *Amer. Min.*, 28 (1943), p. 571.

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crystallization indicate that once a pegmatitic residuum has accumulated, it has remained static until crystallization has been completed. The Cornish pegmatite illustrated in Fig. 77 shows the

-10 cm.-

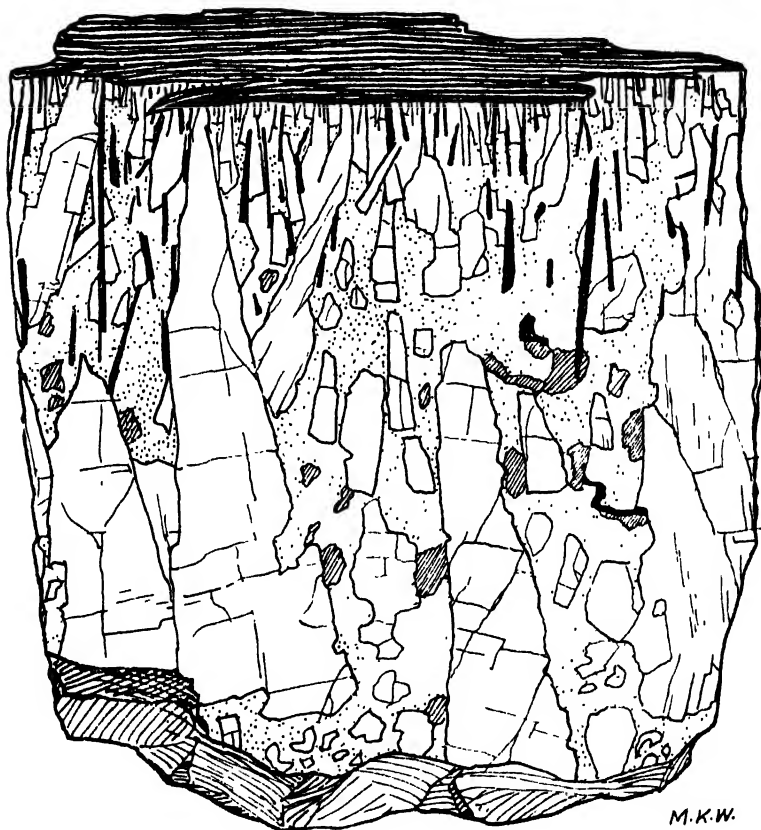


FIG. 77

Diagram of block of granite-pegmatite showing the arrangement of crystals developed perpendicular to the roof of the sill, Porthleven, Cornwall. Tourmaline crystals, black; mica, shaded; quartz, stippled; and perthite, blank. Roof rock is tourmalinized shale.

growth of crystals, including thin acicular tourmalines, perpendicular to the roof of the sill-like intrusions from which the specimen was collected, and it is difficult to imagine the growth of such crystals in anything other than an undisturbed medium.

The *paragenesis* or order of crystallization of the several constituents has been worked out in full detail for a large number of

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pegmatites. It is usually found that intergrown quartz and alkali-feldspar (graphic granite) head the list. The subsequent course of mineral deposition appears to be controlled by the gradual passage of solutions of varying composition through the "pegmatitic thoroughfare." These from time to time are liable to effect changes in composition and textural relationships between the original components. In this way way the albite (clevelandite) of the complex

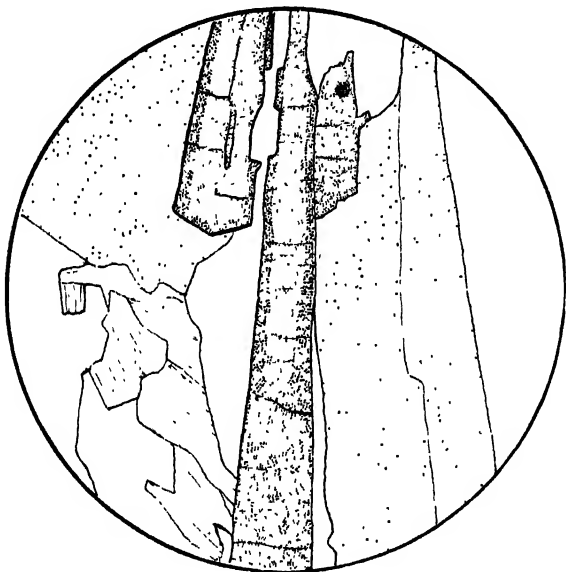


FIG. 78

Granite-pegmatite from pegmatite-aplite sill, an offshoot of the Godolphin granite, Porthleven, Cornwall.

From the upper part of the specimen shown in Fig. 77.

Zoned tourmalines (heavy stipple); orthoclase (light stipple); muscovite (lined); quartz (blank).

pegmatites is produced at the expense of the earlier microcline-microperthite.¹ Striking replacement textures between these two minerals are so frequently observed that the mechanism of this type of albitization is beyond question.

When the migrating solutions dissolve previously formed minerals faster than they deposit new ones, open cavities or vugs may result. These are often lined with fine crystals of smoky quartz, adularia, clevelandite, etc., and with hydrothermal incrustations. The quartz crystals in these cavities are in some cases continuous with the

¹ Schaller, W., "The Genesis of Lithium Pegmatites," *Amer. Journ. Sci.*, 10 (1925), p. 269.

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quartz rods in surrounding graphic granite. The quartz crystals which grew freely into the cavities are of the alpha-type, while the intergrown rods show signs of having inverted from the beta-quartz type. In other words, during the growth of the quartz rods towards the vugs, the temperature fell below the inversion point at which high- is converted into low-temperature quartz: *i.e.* at about 575° C. Thus for once we can make use of a reasonably accurate point on a geological thermometer, and state that pegmatite containing graphic granite normally commences to crystallize not far above 575° C., say at 600° C.

One of the most distinctive mass-characters of many pegmatites is a more or less pronounced zonal structure within the rock-body, shown most obviously by textural differences, and on closer examination by mineralogical differences between successive zones. Thus mica may be represented by biotite in the outermost zone, giving place inwards to muscovite, and this, in turn, to zinnwaldite, and in the centre, to lepidolite. Such zoning is consistent with the idea of the "pegmatite-thoroughfare" outlined above, and in this specific case indicates progressive enrichment in lithium, maximum concentration occurring in the innermost, latest fraction.

Granite-aplites

These rocks occur as veins, as a rule, only a few inches thick, and although most abundant in the parent granite itself, they sometimes penetrate beyond its boundaries into the adjacent rocks. They are found in association with mica-lamprophyres (as rich in mafic minerals as aplites are poor), to which they are *complementary* in Brögger's sense. Chemically they are characterized by a high silica content and a considerable proportion of alkalis, with iron and magnesia in subordination. To this they owe their light colour—white to buff. In the hand-specimens they present a remarkably even and fine-grained saccharoidal texture, which, under the microscope, is seen to be microgranitic or micrographic. The dominant constituent is usually feldspar—a potassic, sodic or lime-bearing variety, according to the type of granite with which the aplites are associated. The proportion of quartz varies, but in some varieties this mineral predominates, indicating a passage to quartz veins.

Relation of Aplite to Pegmatite

Aplites and pegmatites both occur as veins, dykes or sills in granites or the nearby country rock. Further, both types may be seen in the closest association in composite intrusions which may be pegmatitic in their marginal parts, but aplitic in the centre. Occasionally the two may be interbanded; or a vein dominantly aplitic may contain

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irregular pockets of pegmatite. This intimate association demonstrates that both rock-types represent granitic residua and that both have been generated in essentially the same way. There has been much speculation as to the cause of the striking difference in grain-size between these two residua. Probably aplites represent a "dry" and poorly fluxed fraction of the residual granitic magma, while the fugitive constituents were concentrated in the pegmatitic fraction. Doubtless the latter would be far more mobile than the former. The truth of this supposition is established at Porthleven, Cornwall, where pegmatite occurs in contact with the *under-side* of sedimentary xenoliths suspended in granitic (largely aplitic) sills. Such local pegmatite has obviously originated by the arrest of the upward-streaming volatiles.

There is no doubt that many pegmatites and aplites have originated in the manner outlined above; but in some cases a metasomatic origin is indicated by the field-relations of the rocks. If it is conceded that some granitic-looking rocks have originated by replacement, there are even stronger grounds for believing that the same is true of some pegmatites.¹

PNEUMATOLYSIS

Following the final consolidation of the magma, the fugitive constituents are released, and escaping through joints and other fissures may effect striking changes in the mineral composition of the parent rock. These effects are covered by the term pneumatolysis, which implies that the fugitives are in a gaseous state. This is probably true in some cases, but not in others. Whether they exist as a true gas phase or not, however, they behave essentially as chemically active solutions. Pneumatolytic modifications are most strikingly displayed by granitic rocks (though they are not restricted to the latter), and include (a) greisenizing and (b) tourmalinization. Kaolinization is sometimes included, but is better regarded as hydrothermal alteration.

(a) **Greisenizing.**—Greisen, composed essentially of white mica and quartz, is one of the most distinctive products of this type of activity, and appears to have three different modes of occurrence, depending upon the degree of consolidation and fissuring of the parent rock.

(1) Most commonly greisen is a marginal modification of granite adjacent to quartz- and mineral-veins. The alteration is very localized: normally it extends for a distance of a few inches only from the contact. Usually there is convincing evidence of replacement, such as the pseudomorphing of feldspar crystals by aggregates of white mica. The latter is often a variety containing lithium and

¹ See King, B. C., *Journ. Geol.*, 46 (1948), p. 459.

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fluorine, such as zinnwaldite. Other fluorine-bearing minerals, notably fluorite and topaz, are also commonly present. Purple patches of fluorite give colour to an otherwise light grey or white rock. Topaz is an almost constant accessory, and in some specimens becomes an important constituent, even exceeding the white mica in amount. The end-product of this line of variation is a topaz-quartz rock, the former making up perhaps nine-tenths of the rock, to which the name "topazfels" is commonly, though incorrectly, applied. Granites vary widely in the extent to which they display pneumatolytic effects, and this type of greisenizing is developed extensively in only a few areas, such as the tin-mining districts in the Erzgebirge in Saxony, in Cornwall and in northern Nigeria.

(2) In other occurrences the volatile-rich residuum does not alter the parent granite, but gives rise to veins and thin dykes of white-mica, quartz rock which crystallizes in fissures in the granite. Such occurrences are fundamentally different in nature and origin from the demonstrably metasomatic greisens noted above. They are just as definitely primary igneous rocks as aplites and pegmatites. Thus if the name "greisen" implies a metasomatic origin as well as a particular mineral assemblage, another name is needed for these rocks. Spurr¹ has suggested the name "esmeraldite" for primary quartz, white-mica rocks. A better alternative is to apply the term "greisen" to all such rocks, and to indicate the mode of origin by appropriate qualifiers.

(3) Large bodies of greisen sometimes occur as apophyses and marginal facies of granite masses. They are not related to jointing or fissuring in any way, and are obviously different from the greisens considered under (1) and (2) above. A typical example is afforded by the outcrop at Grainsgill, just north of the main part of the Skiddaw granite, Cumberland. The chemical gains and losses of the greisen as compared with the main granite are different in this case from those which usually apply.² There is usually a considerable increase in H_2O , F and probably Al, while Na may decrease to zero. Other components may show small but haphazard changes. When the analyses of Skiddaw granite and Grainsgill greisen are compared, there is found to be an increase in SiO_2 in the latter, which is too great to be accounted for by hydrothermal alteration.³ A. Harker⁴ suggested that the greisen crystallized from a particularly acid

¹ "The S. Klondyke District, Esmeralda County, Nevada," *Econ. Geol.*, 1 (1906), p. 382.

² A valuable summary of the gains and losses involved in greisenizing is given by G. J. Williams: "A Granite-schist Contact in Stewart Island, New Zealand," *Quar. Journ. Geol. Soc.*, 90 (1934), p. 348.

³ Hitchen, C. S., "The Skiddaw Granite and its Residual Products," *Quar. Journ. Geol. Soc.*, 90 (1934), p. 158.

⁴ "Carrock Fell granophyre and Grainsgill greisen," *Quar. Journ. Geol. Soc.*, 51 (1895), p. 143.

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magma-fraction which was separated from the main granite and driven northwards by filter-press action. As the greisen yields evidence of replacement of the original feldspars, however, it is probable that two processes have been involved: firstly differentiation on the lines suggested by Harker, and secondly, deuteric alteration by an active residuum. Retention of the late-stage solutions in such large volumes must depend on such geological "accidents" as the shape of the differentiated body—ideally an apophysis of the main mass—and the degree of fissuring of the surrounding rocks, which must affect the escape of the volatiles.

This three-fold division of greisens and greisen-like rocks into (1) primary, (2) metasomatic and probably pneumatolytic and (3) deuteric, is paralleled among granites affected by tourmalinization.

(b) **Tourmalinization.**—Tourmaline, like muscovite, appears as a normal constituent of some granites of the more acid and alkali-rich types, as in the Carnmenellis mass in Cornwall. A particular concentration of boron produces a brown iron-rich tourmaline in place of biotite, so that the two minerals are very seldom seen together. With increasing flux-concentration tourmaline increases at the expense of other constituents of the granite; in particular, the feldspar is progressively eliminated. *Luxullianite*¹ represents an arrested stage of tourmalinization, in which some brick-red feldspar has survived, though the outlines of the crystals are much corroded. Between them quartz and finely acicular black tourmaline are much in evidence (Fig. 79).

Under the microscope luxullianite is a beautiful rock. The black tourmaline (schorl) occurs in the form of delicate needle-like crystals radially disposed and often clustered around corroded relics of earlier "massive" brown tourmaline. The tourmaline needles freely penetrate into secondary quartz, a mosaic of which forms the general background of the sections. The feldspars have been deeply embayed and show progressive replacement chiefly by quartz.

A second stage of tourmalinization is represented by rare tourmaline-quartz rocks in which none of the original components save quartz has survived; but although completely replaced, the shapes of the feldspars can still be detected, though their substance is chiefly quartz-mosaic.

In other instances tourmaline occurs in imperfect stouter prisms of random orientation, embedded in coarse quartz mosaic, giving a black-and-white rock of distinctive appearance. A well-known example of this type is that of Roche Rock, Cornwall, which forms an isolated outcrop with no visible connection with the nearby main

¹ Wells, M. K., "A Contribution to the Study of Luxullianite," *Min. Mag.*, 27 (1946), p. 186.

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granite intrusion. There is, indeed, no evidence, microscopic or otherwise, that this rock ever was a normal granite: it may well represent a magma-fraction drastically enriched in "fugitives" and analogous with the "primary greisen" noted above. Apart from such

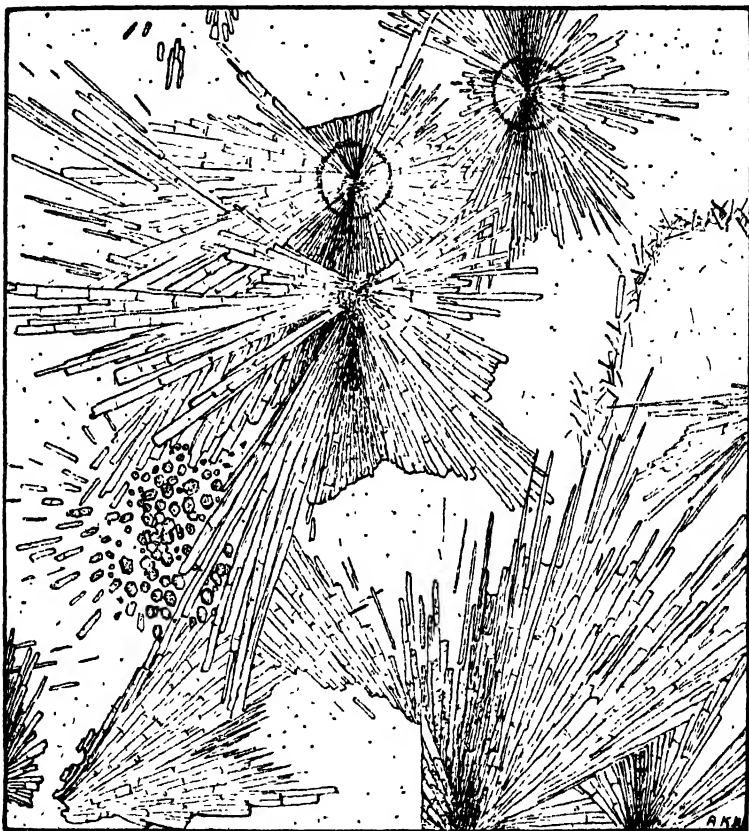


FIG. 79

Section of luxullianite (the type-rock), Luxulyan, Cornwall.

Tourmaline in radial aggregates; quartz of two periods, some pre-tourmaline, some secondary after feldspar. The boundary of a crystal of the latter, completely silicified, is outlined with acicular tourmaline on right, near top of section. Longest tourmaline crystals are 2 mm.

occurrences, however, the same mineral assemblage, sometimes with cassiterite in addition, occurs in the form of primary veins.

Finally, it may be noted that the activities of the fugitive constituents are not limited by the boundaries of the parent granite, for the surrounding rocks are sometimes extensively tourmalinized.

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(c) **Kaolinization.**¹—The partial alteration of feldspars into an exceedingly fine aggregate of flaky minerals of the kaolinite type is a ubiquitous phenomenon in granitic rocks. Such kaolinite is usually accompanied by sericite, and together these two minerals are the chief cause of the opacity of orthoclase in the hand-specimen and its cloudiness in thin section. Occasionally kaolinization is so complete that the whole rock is reduced to a thoroughly rotted condition, becoming friable and so soft that the rock can be easily dug away, or even removed by a high pressure jet of water. In this condition the rock is of great economic importance as **china clay**, perhaps the most valuable raw material in the ceramic industry. Of the original minerals of the granite, only the quartz survives kaolinization. Large masses of the St. Austell granite² in Cornwall have been modified in this way, giving china clay in association with other late-stage alteration products such as quartz-tourmaline rock and **china stone**. The latter rock is comparable with luxullianite in a sense, as it represents an arrested stage of alteration—this time of kaolinization—and exhibits a wide variety of replacement textures. Two feldspars are present: strongly kaolinized orthoclase, often microperthitic, and relatively fresh euhedral albite. The orthoclase may be extensively veined by secondary quartz and by fluorite. The latter is an essential component in more than the usual petrographical sense, for it gives the rock its fluxing properties in the glazing of china. It occurs in a variety of forms, either replacing feldspar or penetrating along the cleavage planes of muscovite as shown in Fig. 80 (top right and left).

The chief agent of kaolinization is probably high-temperature H_2O which attacks the orthoclase ($KAlSi_3O_8$), etc., to form kaolinite, $Al_2Si_2O_5(OH)_4$, and releases a certain amount of potassium which may combine with CO_2 to form a carbonate and thus be removed as a soluble component; or the potassium may combine with other components to form sericite which is so commonly associated with the kaolinite. Carbon-dioxide may thus play a part in these reactions, but it is uncertain how important a part.

MICROGRANITES, MICRO-ADAMELLITES AND MICRO-GRANODIORITES

In the view of the practical difficulty of estimating the relative amounts of potassic and sodic feldspars (except by calculation of the norm), it is necessary to be satisfied with generic, rather than specific, identification. Actually there is little point in trying to

¹ "Handbook to the Collection of Kaolin, etc.," *Mém. Geol. Surv.*, 1914.

² Exley, C. S., "Magmatic differentiation and alteration in the St. Austell granite," *Quar. Journ. Geol. Soc.*, 114 (1959), 197-230.

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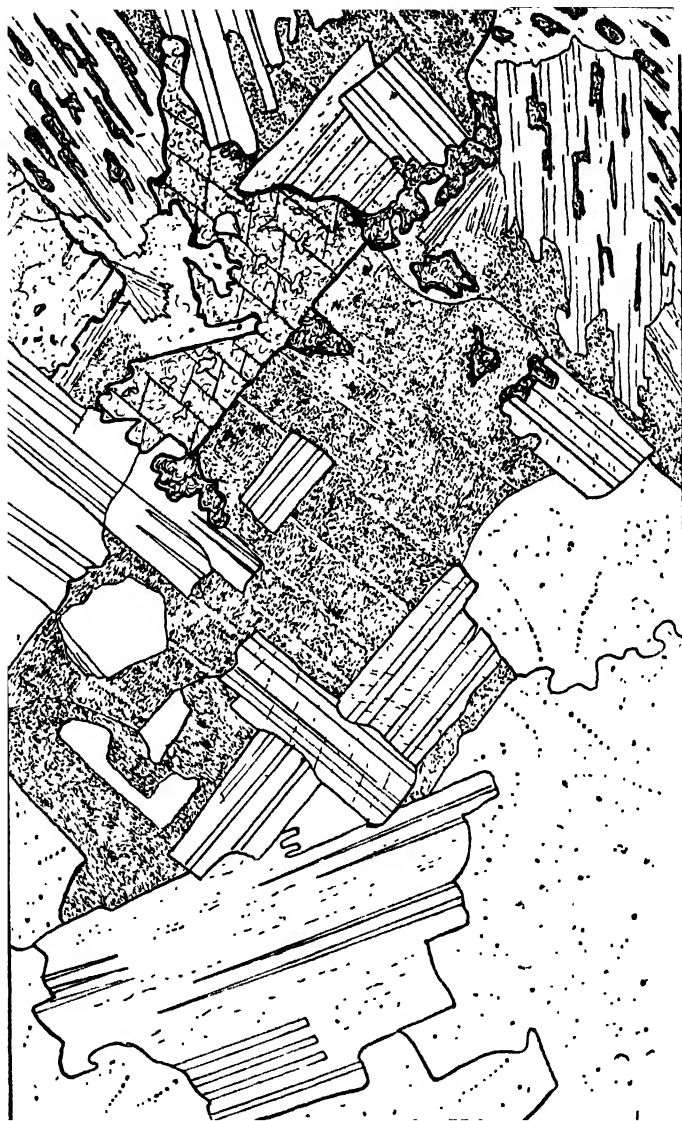


FIG. 80

Cornish "china stone." Orthoclase (stippled) is strongly kaolinized, while albite is relatively fresh. Quartz contains numerous inclusions. Fluorite is conspicuous in muscovite, and in larger, colourless interstitial patches, showing octahedral cleavage. One topaz crystal top left of the section, about 1 inch down.

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specify that a given microgranite is either potassic or sodic—most are potassic-sodic or sodi-potassic. The feature that attracts more attention than the composition of the feldspar is the nature of the coloured silicate(s) present.

In some microgranites biotite occurs; but it is evident that these dyke-rocks are nearer to rhyolites than to granites so that pyroxene—usually a light coloured diopsidic augite—takes precedence over mica. Hedenbergite, accompanied sometimes by iron-rich olivine, occurs in microgranites from northern Nigeria; while ferrohortonolite occurs in a Tertiary granophyre (Meall Dearg, Skye). In more strongly sodic microgranites such NaFe-rich pyroxenes and amphiboles as aegirine and riebeckite occur—they have already been noted as occurring characteristically in granites of sodic types. It is customary in naming these rocks to hyphenate the mafic mineral-name on to the term “microgranite”: thus aegirine-microgranite and riebeckite-microgranite are informative names. The former is still disguised under the term *gorudite* (Brögger, 1894): from the details of its composition, SiO_2 , 76 per cent, orthoclase-microperthite (with large excess of K_2O over Na_2O), aegirine and, of course, quartz, it is evident that *gorudite* is synonymous with aegirine-microgranite. Nothing is gained by retaining this rock-name.

As regards texture, microgranites grade with diminishing grain-size into rhyolites, and with increasing grain-size into granites—the limits in both directions are arbitrary; the textures are intermediate between, and grade into those displayed by rhyolites and granites respectively. Three different textures claim our attention: (1) porphyritic, (2) aphyric (the clumsy term “non-porphyritic” is often preferred, unfortunately); and (3) micrographic.

(1) **Porphyritic microgranites, micro-adamellites and micro-granodiorites** are megascopically the most distinctive members of the family (Fig. 81). They have been widely termed “quartz-porphyries” or “granite-porphyries” though the terms favoured here are less liable to misinterpretation than the other two. The phenocrysts (first generation crystals) are believed to have formed in depth, about relatively few, widely spaced centres of crystallization in an environment where growth was uninhibited and perfect crystals were developed, notably the feldspars which are euhedral and plane-faced. Quartz is the high (β) form, of distinctive dihexagonal bipyramidal shape; the crystals are both fewer and smaller than the feldspars. A handsome British example forms a dyke striking across the beach at Prah Sands in South Cornwall and known as the “Prah elvan.” It is noteworthy for the fluxional disposition of the abundant feldspar phenocrysts it contains. The matrix of these rocks is visibly granular and is seen in thin section to be a microcrystalline granular mosaic of quartz and alkali-feldspar.

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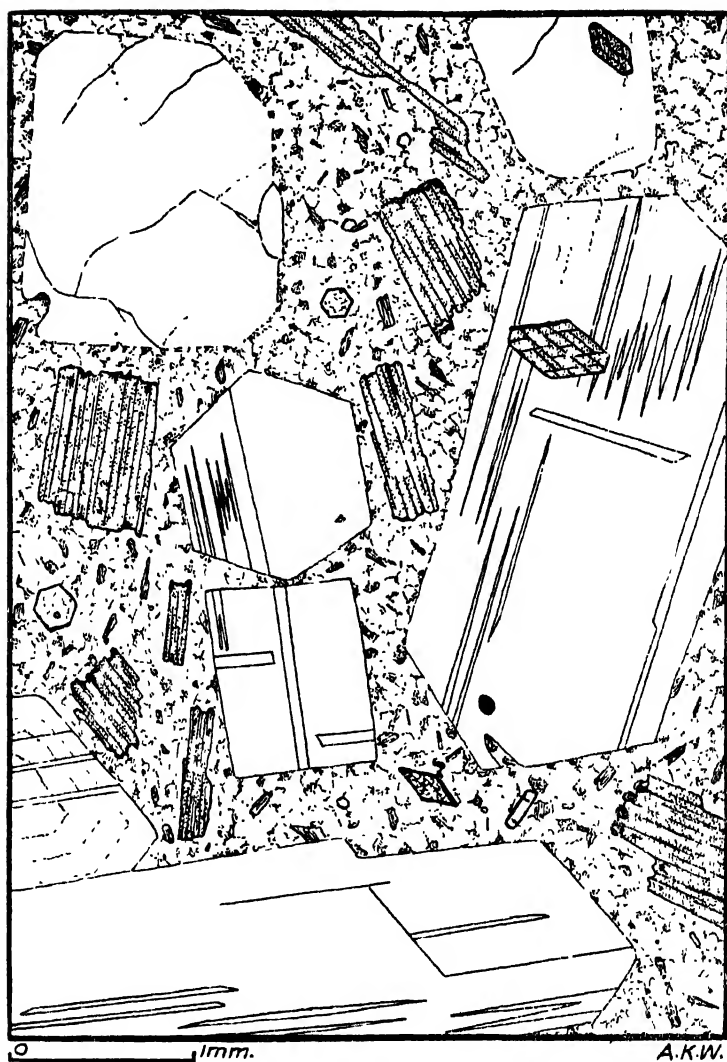


FIG. 81

Porphyritic micro-granodiorite, south of North Bend, British Columbia.

Phenocrysts of quartz, zoned plagioclase, common hornblende and biotite are embedded in a microcrystalline groundmass of the same minerals. Apatite and sphene are also shown. The twinning is indicated in the plagioclase.

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(2) **Aphyric microgranites**, etc., resemble the previous rocks in composition and mode of occurrence; but as the qualifier implies, there are no porphyritic crystals: in a sense they are all ground-mass. Texturally and, up to a point, mineralogically too, these rocks are close to granite-aplites but differ by containing appreciably more coloured mineral: aplite represents only a felsic residue from granite, but an aphyric microgranite represents *all* the granite, mafic as well as felsic components.

(3) The most interesting texture is the **micrographic** which has several features in common with the graphic, considered above. "Micrographic microgranite" is a ponderous term and **granophyre** (Rosenbusch, 1872) is a more acceptable name for this group of rocks. Most granophyres are porphyritic and, in general, are close to porphyritic microgranites: in fact it is only the groundmass texture which distinguishes them. This consists of intergrown quartz and alkali-feldspar, the former occurring as isolated "hieroglyphs" like those seen in the coarser-grained graphic granites and pegmatites, or sometimes as "fibres" which have grown away at right angles to the faces of the feldspar phenocrysts and which expand in girth distally. The hieroglyphs or fibres are in parallel optical orientation within each area of micropegmatite and therefore extinguish simultaneously (Fig. 82). The micropegmatite patches may be roughly rounded, and within each, the quartz hieroglyphs are arranged centrically. This type of structure grades into the spherulitic as the quartz individuals become finer, thinner and more regular in their radial disposition.

With regard to their nature and origin all granophyres were at one time accepted as being magmatic rocks of essentially granitic composition; but certain granophyres are now believed to be metasomatic in origin. The evidence seems to be convincing in the case of certain granophyres which are marginal to Karroo dolerites, and which appear to have been formed by the metasomatism of the sediments into which the doleritic magma was intruded.¹ There is nothing inherently improbable in this; but we think it is unwise to conclude on this evidence that all granophyres have originated in the same way—a view that is apparently held in some quarters.² It may be appropriate, therefore, to restate the case for believing that most granophyres were formed by direct crystallization of a melt, and must therefore be regarded as magmatic. The evidence concerns the composition and textures of the rocks, their field relations and affinities with other rocks, notably certain rhyolites.

¹ Walker, F., and Poldervaart, A., "Karoo dolerites of the Union of S. Africa," *Bull. Geol. Soc. S. Africa*, 60 (1949), p. 591.

² See, e.g., Holmes, A., in discussion on Hughes, C. J., "The Southern Mountains igneous complex, Isle of Rhum," *Quar. Journ. Geol. Soc.*, 116 (1960), p. 111.

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FIG. 82

Aphyric granophyre, Fjardardal, Iceland.
Orthoclase is shown stippled, quartz black and white.

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First as regards mineral composition: the proportion of quartz to alkali-feldspar corresponds to the lowest-temperature range of compositions established experimentally for the synthetic "granite" system, quartz-orthoclase-albite. Characteristic mafic minerals include pyroxenes and olivines of magmatic types, including hedenbergite ($\text{Ca}_{42}\text{Mg}_7\text{Fe}_{51}$) and ferrohortonolite (Fa_{85}) occurring in the Meall Dearg granophyre in Skye: the same minerals are found in high-level late fractions forming part of the Skaergaard complex, also in rhyolitic glasses including pitchstones. As regards texture, the fact that it can be closely matched among metallurgical textures produced during crystallization of eutectic mixtures has long been regarded as significant. So, to us, is the fact that the textures observable in the mesostasis of quartz-dolerites are identical with those seen in some granophyres. The textures in the quartz-dolerites are magmatic and it would be extraordinary if identical textures in granophyres were produced in some other way.

Finally, granophyres occur in the so-called volcanic association. In composition they closely match certain rhyolites; they have the same mode of occurrence as various microgranites and pitchstones; they show the same close association with basic (basaltic) rocks in various Acid/Basic associations, including composite dykes and sills, while the recently described granophyre pipes occurring in doleritic and gabbroic rocks on Slieve Gullion and elsewhere are particularly significant in this respect.¹

We would emphasize one point. We are satisfied from the above evidence that the majority of granophyres are magmatic. How the magma was formed is another matter. Selective melting of material of the appropriate composition by basic magma at a temperature of 1,100 to 1,200° C. is well within the realms of possibility; and we would class arkose, rhyolitic "ash" and an older granodiorite among the materials "of appropriate composition."

One of the most recently described granophyres is that which forms the scarp feature of Cader Idris in Merionethshire, North Wales.² Its chilled facies is indistinguishable from certain rhyolitic lavas with which it is closely associated. It is interesting to note that this granophyre has *caused* metasomatism and mobilization of the sediments into which it was intruded, under a cover perhaps 1,500 feet thick. This illustrates how difficult it may be to sort out cause and effect.

Microgranites are widely distributed. Types corresponding in composition with the alkali-granites are associated with, and occur

¹ Elwell, R. D., and others, "Granophyre and hybrid pipes . . . of Slieve Gullion," *Journ. Geol.*, **66** (1958), pp. 57-71.

² Davies, R. G., "The Cader Idris Granophyre and its associated rocks," *Quar. Journ. Geol. Soc.*, **115** (1959), pp. 189-216.

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as offshoots from, major intrusions of this nature. Thus the coarser-grained dyke-rocks, locally termed "elvans," found in the neighbourhood of, or actually cutting the Cornish and Dartmoor granites, are potassic or sodi-potassic microgranites. Those with adamellitic and granodioritic affinities tend to occur in association with, for example, the Caledonian "granites" of these types in Scotland. Many of the minor intrusions in the English Lake District are microgranites: the St. John's Vale intrusion, characterized by phenocrysts of orthoclase, quartz and small red garnets, is a porphyritic potassic microgranite. Among types which are correspondingly sodic we may note **riebeckite microgranite**, which occurs at Ailsa Craig in the Firth of Clyde, on Holy Island off the coast of Arran and at Mynydd Mawr in North Wales (Figs. 83 and 84). Although closely similar in mineral composition, the two examples figured differ in textural detail. The micropoikilitic habit of the riebeckite, sometimes giving it a "mossy" appearance, is characteristic.

It may be noted that the Ailsa Craig rock has been called "riebeckite-aplite." But the term "aplite" implies not only a peculiar mineral composition and a distinctive texture, but a special mode of origin. Unless there are strong grounds for believing that the rock in question has originated in the manner described above under "aprites" the term aplite should definitely not be used.

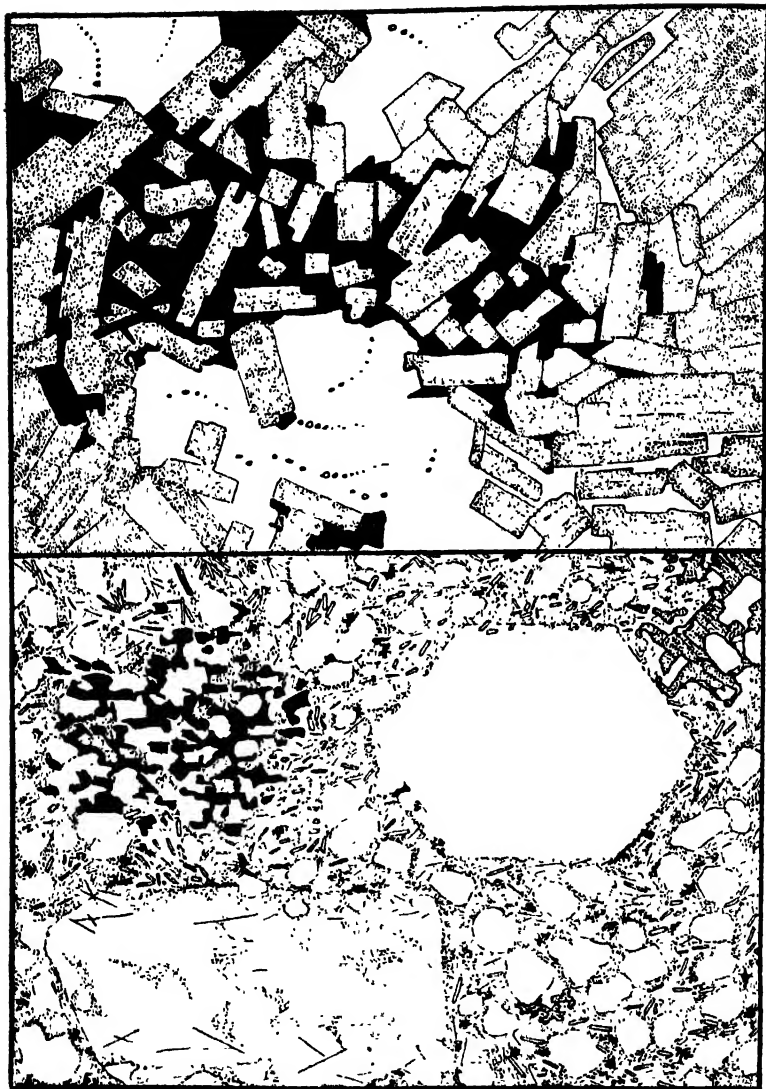
Aegirine-microgranite occurs among the dyke rocks connected with the alkali complexes in Sutherlandshire, for example, Cnoc-na-Sroine in Assynt.

One of the best known British examples of granophyre is that which occurs at Carrock Fell: the intergrowth in this case is relatively coarse. The rock forming the Armboth dyke in the Lake District is a handsome porphyritic variant with phenocrysts of quartz, bright red orthoclase and small garnets set in a dun-coloured graphic ground-mass. The prominent Fort Regent mass lying on the outskirts of St. Helier, Jersey, is only in part medium-grained, but is beautifully micrographic in places. The massive columnar sill, in the main a rather fine-grained granophyre which forms the formidable scarp feature of Cader Idris in North Wales, has already been mentioned.

THE RHYOLITES

If for the sake of simplicity one collective name is required to include all the Acid igneous rocks of fine grain, it must be rhyolite, which thus has the same status as "granite" for coarse-, and "microgranite" for the medium-grained rocks. This is one of the older rock names, and was used by von Richthofen (1860) on account of the flow banding frequently exhibited by these rocks. The name would serve

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A K W.

FIG. 83 (*above*).—Riebeckite microgranite, Ailsa Craig, Firth of Clyde, Scotland. Alkali-feldspar embedded in poikilitic riebeckite (shown black) and quartz.

FIG. 84 (*below*).—Riebeckite-acmite-microgranite, Mynydd Mawr, Carnarvonshire. Phenocrysts of β -quartz and alkali-feldspar in microcrystalline ground-mass, with acicular prisms of acmite, and micropoikilitic riebeckite, top right and near top left—black (in position of maximum absorption).

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a more useful purpose if it were restricted to those fine-grained rocks which are strictly analogous with alkali-granites. "Rhyolite," thus used, is of the same status as rhyodacite and dacite. Liparite (Roth 1860), from well-known occurrences in the Lipari Islands, is synonymous with rhyolite.

The upper limit of grain size is 0.05 mm., this having reference to the diameters of quartz or feldspar grains in the groundmass. Grains of this size are, of course, irresolvable with the naked eye, so for practical purposes the dividing line between microgranites and rhyolites can be placed at the limit of unaided vision for the groundmass components. No limit of size is laid down concerning the phenocrysts which are present in many specimens.

In attempting to subdivide this large class, we try to apply the same criteria as for the granitic rocks, and recognize three main categories based on the feldspar contents. It must be realized, however, that in many cases on account of fineness of grain or the presence of natural glass, it is impossible to apply any but a general name: specific identification along these lines is impossible. With an analysis available, the task is much simplified, of course.

ALKALI RHYOLITES (*RHYOLITES sensu stricto*)

In the scheme of classification these are exactly analogous to the alkali-granites and microgranites, and, like them, may be further subdivided into (a) potassic and (b) sodic types, in favourable circumstances.

Potassic Rhyolites.—In these the feldspar consists of the high-temperature form of orthoclase—sanidine, often in the form of glassy clear phenocrysts and/or as minute granules or microlites in the groundmass. Sodic plagioclase is not excluded, but must be subordinate to the sanidine. In all rhyolites free silica may be present not only as inverted β -quartz,¹ but also as tridymite and even cristobalite. Of these β -quartz is by far the commonest, and the only obvious form of silica even under expert examination. It occurs as well-formed bipyramids, which, however, may show any degree of magmatic corrosion, and in extreme cases may be reduced to shapeless wrecks of the original crystals (Fig. 85). By comparison with granite one would expect the micas and hornblende to be the most commonly occurring coloured silicates; but actually although biotite is common, hornblende is rare in rhyolites, pyroxene is widely distributed. This is consistent with the higher temperature

¹ Readers may be reminded that all quartz in igneous rocks, in rhyolites as well as in granites, is "low" quartz. That occurring in the rocks under discussion was precipitated as "high" quartz but subsequently inverted into "low" quartz during cooling.

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at which consolidation took place in the case of rhyolite, and the boiling off, instead of retention, of volatiles in lava flows doubtless also favoured the crystallization of the anhydrous pyroxene. The biotite is often very strongly coloured, so that basal sections appear almost black. The pyroxene is often nearly colourless—actually a very watery light green—in thin section, and is a diopsidic augite.

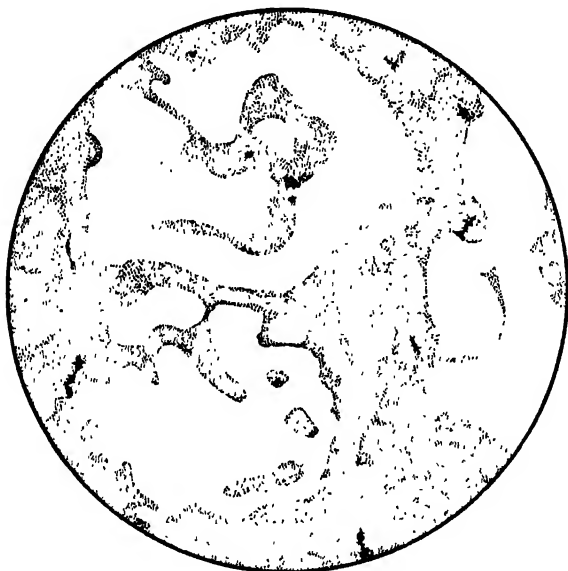


FIG. 85

Devitrified porphyritic pitchstone, Petit Portelet, Jersey, C.I. Highly corroded phenocrysts of quartz in "felsitic" (originally glassy) groundmass.

Sodic Rhyolites.—Like sodic granites these rocks may be very distinctive, largely by reason of the coloured silicates they contain, which may be the same as those already described in the corresponding granites and microgranites: *viz.*, aegirine among the pyroxenes and riebeckite among the amphiboles being characteristic (Fig. 86). The coloured silicates are accompanied by quartz, of course, and ideally by albite, though in very many instances the feldspar, as in the corresponding coarse-grained rocks, is sodi-potassic, often recorded as anorthoclase which would unmix, presumably to antiperthite, in a plutonic environment.

From the facts of their mineral composition such riebeckite- and aegirine-rhyolites are evidently comagmatic with the corresponding granites and dyke-rocks with which they are closely associated in the field, for example in the alkali-complexes in Nigeria (Fig. 86).

ROCKS CONTAINING ESSENTIAL QUARTZ

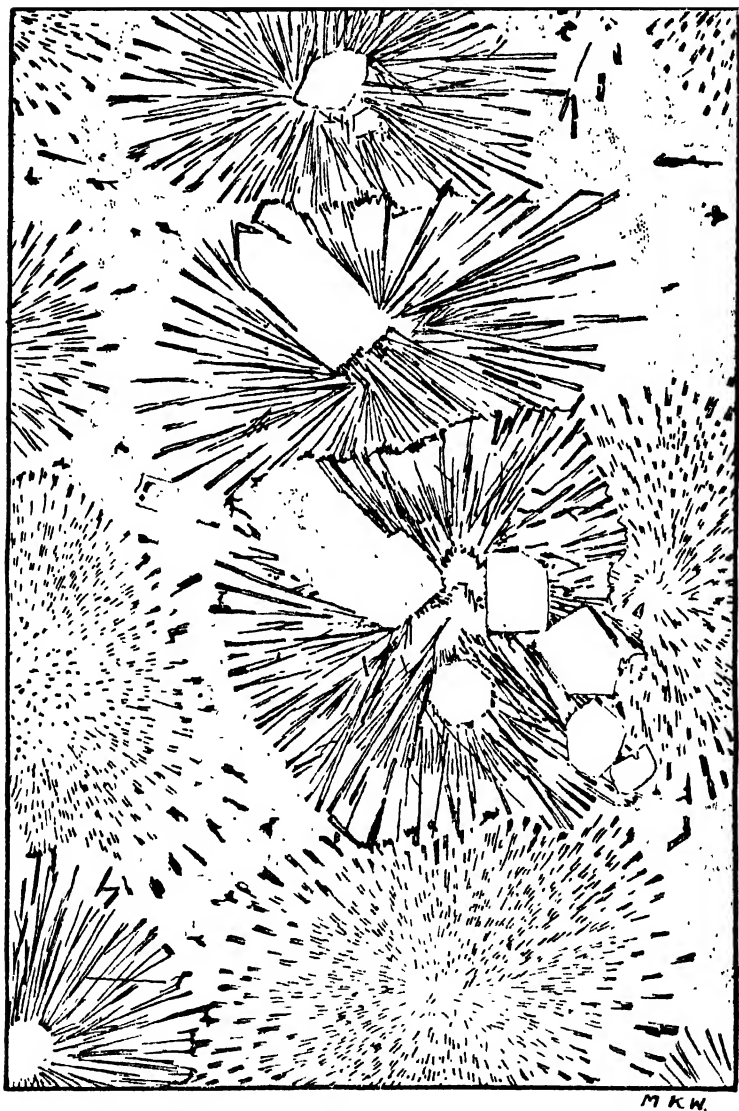


FIG. 86

Spherulitic riebeckite-rhyolite, between Rishi and Ayaka Leru,
northern Nigeria.

Spherulites of riebeckite enclosing phenocrysts of "high" quartz and sodipotassic
feldspar in a devitrified (felsitic) matrix.

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The type illustrated is a very unusual one as the mineral involved in the spherulites is riebeckite which adventitiously encloses small phenocrysts of sodipotassic feldspar and inverted beta-quartz. The groundmass is a devitrified glass, represented by a felsitic aggregate of quartz and feldspar. Some types of sodic rhyolite have been given specific names. Thus "quartz-pantellerite," so named by Washington from occurrences on the island of Pantelleria, near Sicily, contains in order of abundance, "soda-microcline" (about $\text{Ab}_{50}\text{Or}_{50}$), quartz, acgirine, or aegirine-augite, and the distinctive and rare amphibole, cossyrite. "Quartz-keratophyre" has been used in several senses, but the practice in this country is to apply it to sodic rhyolites characterized by the presence of albite or albite-rich feldspar, associated with a little coloured silicate, the original nature of which has become obscure through alteration: it is represented by uninformative patches of chlorite. Lavas and texturally similar minor intrusives of this type occur among the Ordovician and Devonian eruptive rocks of North Wales and Devon and Cornwall respectively. Quartz-keratophyre is a type of soda-rhyolite, and a member of the spilitic suite.

ANALYSES OF ALKALI-RHYOLITES

	I Potassic rhyolite.	II. Pitchstone, Arran.	III. Sodi-potassic rhyolite- obsidian.	IV Sodic- rhyolite (Pantellerite).
SiO_2	73.76	73.20	75.52	68.63
Al_2O_3	11.98	10.75	14.11	10.30
Fe_2O_3	1.14	0.95	1.74	5.60
FeO	2.40	1.02	0.08	2.61
MgO	0.76	0.15	0.10	0.37
CaO	0.32	0.76	0.78	1.07
Na_2O	0.53	3.78	3.92	6.14
K_2O	7.38	4.20	3.63	4.17
H_2O	1.75	4.70	0.39	0.53
TiO_2	0.12	0.16	—	0.35
Other constituents	0.34	0.61	0.11	0.23
	100.48	100.28	100.38	100.00

I. Potassic rhyolite. Cwm Caregog, Snowdon (Anal. R. J. C. Fabry), *Quar. Journ. Geol. Soc.* (1927), p. 368.

II. Sodi-potassic pitchstone, Arran, Scotland (Anal. E. G. Radley), *Geol. of Arran, Mem. Geol. Surv.* (1928).

III. Obsidian, Obsidian Cliff, Yellowstone National Park, U.S.A. (Anal. Whitfield).

IV. Sodic-rhyolite (Pantelleria type). Average of twelve pantellerites quoted from Daly (1933), p. 10.

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RHYODACITES (TOSCANITES)

The rocks in the fine-grain category corresponding with adamellites are known by two names: the older is toscanite (Washington, 1897) derived from the Italian name for Tuscany, whence the original rocks so named were described. The alternative name is more informative: **rhyodacite** conveys the correct impression of the essential feature of these rocks—the sharing of the characteristics of rhyolites and dacites. Thus in rhyodacites two kinds of feldspar, an alkali type characteristic of rhyolites, and a sodic plagioclase of the kind occurring in dacites, are of equal status. The term is comparable with trachyandesite, which occupies an analogous position among the “Intermediate” lavas. Most rhyodacites are porphyritic and all three felsic constituents may occur as phenocrysts, though commonly the latter include inverted beta-quartz and plagioclase, while the alkali-feldspar, sanidine, forms the major part of the groundmass.

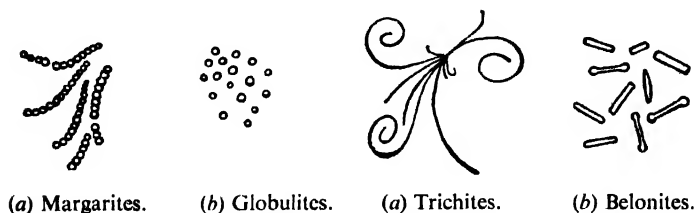
DACITES

Dacites may be defined as occupying the position among the fine-grained igneous rocks of granodiorites among those of coarse-grain size. In terms of mineral contents, this implies a marked predominance of plagioclase over alkali-feldspar, to the extent of at least two-thirds of the total feldspar. Theoretically dacites, like other types of acid lavas, may contain phenocrysts, or may be free from them. In practice the name is seldom applied unless phenocrysts of both quartz and plagioclase are present. Clearly only a portion of the total free silica will be visible as phenocrystic quartz—much of it will be retained in the groundmass, and a chemical check is desirable in doubtful cases. Further, while it was practicable to differentiate between granodiorite and quartz-diorite, it is difficult when dealing with their fine-grained equivalents, even if thin sections are available.

Textural range of the rhyolites.—The rhyolites which capture the popular imagination are undoubtedly the vitreous types. Although the proportion of glass to crystalline mineral in rhyolites is infinitely variable, those types richest in glass are termed obsidian, pitchstone or pumice. At its best **obsidian** is a pure natural glass of rhyolitic (or granitic) composition, black in colour, naturally vitreous in appearance, and usually exhibiting a striking conchoidal fracture. Obsidian may be easily trimmed up to any desired shape—to a point or sharp cutting edge, for example; therefore this rock was largely

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used by primitive peoples in the manufacture of stone implements and weapons. Although obsidian is the commonest of the natural glasses, it is not widely distributed, the occurrences at Obsidian Cliff in the Yellowstone Park, at Mount Hecla in Iceland, and in the Lipari Isles, being the best known. In thin section obsidian is colourless and isotropic. It is rarely completely devoid of crystalline material, which may occur as minute scattered crystallites (Fig. 87),



* FIG. 87

Some types of crystallites.

or spherulites, either isolated, in irregular groups or trains, or in definite bands (spherulitic obsidian). Although the proportion of glass in **pitchstone** is very high, there is much more crystalline material than in obsidian. It may take the form of phenocrysts, of any of the minerals appropriate to rhyolites— β -quartz, sanidine, oligoclase, light green pyroxene are all common—or as microlites. The latter have been identified in some cases as feldspar, in others as pyroxene, presumably of the same type as that occurring as phenocrysts. The microlites may appear rod-like in thin section, but in some cases they form stellate or feathery groups (see Fig. 90), or may resemble fern fronds. In these glassy rhyolites of both types flow structure is common: it is shown sometimes by the parallel alignment of the microlites which may sweep round phenocrysts like eddies in a stream, and convey a vivid impression of the viscous flow of the magma. In other cases the glass itself is colour-banded (Fig. 88), or layers of vitreous and non-vitreous lava may alternate, doubtless resulting from a degree of heterogeneity in the lava. Perlitic structure, resulting from tension set up by contraction during cooling, occurs in both obsidian and pitchstone (Fig. 89). These two terms are used rather loosely: in the opinion of some, the distinction between them should rest on the different degree of crystallinity, as noted above; but in the opinion of others, the difference in water content is more significant. In obsidian the amount is small—usually less than 1 per cent—while in pitchstone it may rise as high as 10 per cent. Johanssen has advocated drawing the line at 4 per cent of water, but although this would make for

ROCKS CONTAINING ESSENTIAL QUARTZ



FIG. 88

Porphyritic pitchstone, Sgùrr of Eigg, Hebrides.

The corroded phenocrysts are of β -quartz and alkali-feldspar ($\text{Or}_{42} \text{Ab}_{52} \text{An}_6$); they have been chipped and the fragments scattered by the flow, indicated by the banding.

accuracy, it is a drawback being unable to name a specimen until an analysis is available.

The common occurrence of glassy rocks among the rhyolites is the natural consequence of the composition of the magma from which they were formed. The chief components, alkali-feldspar and quartz, are present in nearly eutectic proportions. Such a mixture remains



FIG. 89
Perlitic obsidian.



FIG. 90
Pitchstone, Corriegills, Arran, with pyroxene crystallites
surrounded by zones of clear glass.

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in the liquid state longer and at a lower temperature than a magma of different composition. At temperatures near the eutectic, crystallization is inhibited by the high viscosity.¹

As regards mode of occurrence glassy rhyolites, equally with those of other types, may occur both as extrusive flows and intrusive sills and dykes. In Britain famous examples of pitchstones occur among the Tertiary intrusives in the Hebrides, in Arran and Eigg. The Arran pitchstones are well known on account of the interesting textures they exhibit (Fig. 90). Further, some contain iron-olivine, fayalite, in addition to ferro-augite and orthopyroxene, feldspars and quartz phenocrysts (Fig. 91). Most teaching collections contain specimens of the Meissen pitchstones which are reddish in colour, and are mostly lavas. Pitchstone lavas occur also among the Old Red Sandstone volcanics in the Cheviot Hills.

All these glassy rocks tend to change their character spontaneously in the course of time, when they become devitrified. Although evidence of their originally glassy nature may be clearly proved by traces of perlitic structure, for example, examination between crossed polarizers shows them to be completely crystalline—often quite coarsely microcrystalline, the whole rock breaking up into a mosaic of ill-defined grey-polarizing areas bearing no relation whatever to the original structure. This is often called felsitic texture, and the rock itself a **felsite**. The term felsite is a very broad one, applied to compact pinkish or grey dyke-rocks satellitic to granite masses. Some felsites are demonstrably devitrified rhyolitic glasses, but in some the crypto-crystalline (not microcrystalline) groundmass may be original. Many examples occur among the lavas of Uriconian age in Shropshire, and of Ordovician age in North and South Wales.

It was mentioned incidentally above that these rocks may be spherulitic. Ideally the spherulites consist of near-spheres, often about the size of a pea, but varying from smaller than a pin's head to larger than a man's fist. In the case of the smaller ones a section shows the spherulites to consist of radially disposed fibrous crystals, which are not easily identified under the microscope (Fig. 92). In some cases they contain cristobalite, or may consist of it.² In certain rhyolites the whole of the groundmass may be spherulitic: the characteristic radial arrangement of fine microlites may be seen, particularly between crossed polarizers, although on account of mutual interference during growth the individual spherulites are no longer spheres, but they may be recognized by the imperfect black cross that results from the straight extinction of the individual fibres. Interesting devitrified examples occur among the Tertiary

¹ Hawkes, L., "On Rock Glass and the Solid and Liquid States," *Geol. Mag.*, 1930, p. 17.

² Howard, A. D., *Amer. Min.*, 24 (1939), p. 485.

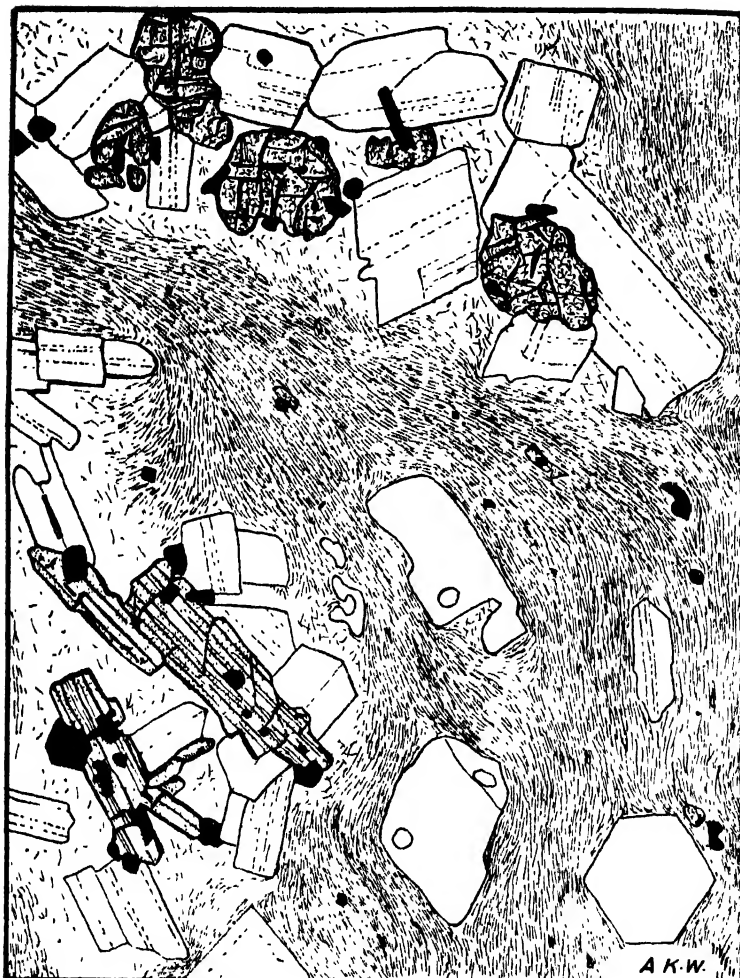


FIG. 91

Fayalite-pitchstone, Glen Shurig, Arran. Fayalite phenocrysts occur in the upper half of the section, a greenish ferroaugite in the lower half. Iron ore, plagioclase (An_{27}) and β -quartz are also present, the latter giving typical hexagonal basal section. The fluxion structure shown by groundmass microlites ("spiculites") is noteworthy.

minor intrusions in Arran: they are usually known as spherulitic felsites. The French name "pyromerides" is still applied to the spherulitic rhyolites of the north coast of Jersey, Channel Isles, which are perhaps unique. The spherulites often show a strongly developed concentric structure, superimposed upon the more normal

ROCKS CONTAINING ESSENTIAL QUARTZ

radial arrangement. In this respect they are allied to the so-called lithophysae—stone bubbles which superficially resemble spherulites but are hollow inside. Lithophysae may reach several centimetres diameter.

With slower cooling rhyolitic magma will not congeal as glass,

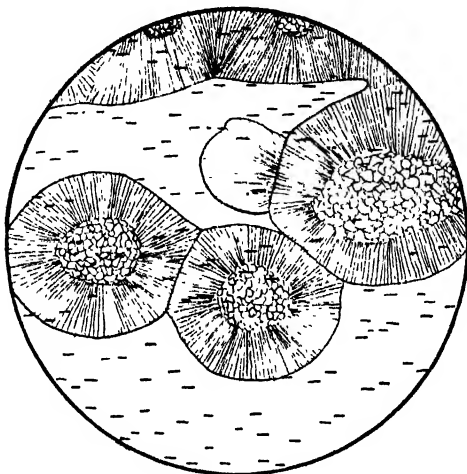


FIG. 92

Spherulitic rhyolite-obsidian, Lipari.

Showing "spiculites" arranged by flow movements prior to the growth of spherulites.

but will become wholly crystalline. There may be difficulty in deciding whether this holocrystalline condition is original or the result of devitrification; but in the absence of relics of the glassy state, referred to above, and if the individual grains are clearly defined, it is almost certainly an original structure. As noted above, rhyolites may be porphyritic to any degree: the phenocrysts may be minute and widely scattered, or they may be relatively large and so closely packed as almost to exclude the groundmass. The name "nevadite" was applied by von Richthofen to a rock of the latter type. It is best, however, to introduce new names for rocks with distinctive mineral composition, rather than some mere textural peculiarity, which can always be covered by appropriate qualifiers. Richness in phenocrysts is not an attribute of one kind of rhyolite, however, and while one lava described as nevadite is a phenocryst-rich potassic rhyolite (with porphyritic sanidine and quartz), another has been called by the same name on account of the abundance of phenocrysts, but as they include oligoclase, biotite and hornblende in addition to sanidine and quartz, we are dealing with a rock in a different category altogether: it is a porphyritic dacite.

PETROLOGY OF THE IGNEOUS ROCKS

Distribution and origin

Within the last decade rhyolites have received considerable attention from petrologists following the recognition of the extraordinary nature of the rocks termed ignimbrites, tuff-lavas and welded tuffs. The last two terms suggest that these rocks are fragmental, pyroclastic and consequent upon volcanic eruptions of explosive type. This is partly true, and for this reason the problem of their nature and origin is considered more fully in the chapter in which pyroclasts are described. It is clear, however, that some ignimbrites are *lavas*, usually but not invariably rhyolitic, which were erupted under unusual conditions. It has been noted that rhyolitic lava is frequently represented by glassy rocks, the failure to crystallize being the direct consequence of the extremely high viscosity of the magma. Normally, therefore, rhyolites occur as flows limited to small areas in the near neighbourhood of the volcano from which they were erupted: indeed many rhyolites just failed to reach the surface and consolidated under a thin cover. Ignimbrite flows by contrast are often very extensive, and it is evident that special conditions of eruption must have obtained. The magma at the time of eruption was effervescing freely and was erupted as a foaming "glowing avalanche" of pumiceous material. Rhyolitic ignimbrites have been recognized in many parts of the world: indeed, they are so widespread that there can be little doubt that this is the normal mode of eruption for magma of this composition. The volcanic gases play a vital part in causing the effervescence and in acting as a lubricant to what would otherwise be an impossibly viscous material.

The important point to note is that in spite of their new name, the rocks involved are rhyolitic, rhyodacitic or dacitic. The more fundamental problem of how and where rhyolitic magma originates is discussed below, under "Origin of the Granitic Rocks."

Rhyolites, rhyodacites and dacites are well represented and widely distributed in ancient and modern volcanic regions. In Britain sodic, potassic and sodipotassic types occur interstratified with the Ordovician rocks, notably in North Wales (the Dolgelley district and Snowdonia) and rather less extensively in the English Lake District. Among the Devonian lavas interbedded with the Old Red Sandstone in Scotland rhyolites occur, for example in the Glencoe area, the Pentland Hills, also in the Cheviot Hills in the Border Country. The Tertiary volcanic episode provides us with striking examples of rhyolitic glasses which are well known: they have been mentioned in the above account and copiously illustrated. The Tardree rhyolite in Northern Ireland is one of the few British rocks known to contain tridymite.

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"GRANITIC rocks" includes not only granites within the whole range of quartz-rich rocks from granodiorites to alkali-granites (true granites), but also their fine-grained equivalents, the rhyolites, and the medium-grained rocks occurring as associated minor intrusions. The lavas and associated dyke-rocks are demonstrably magmatic; so too are the comagmatic granites, and for these assemblages there is no problem of origin: they represent rhyolitic magma cooled under different conditions. The reality of rhyolitic magma is emphasized every time rhyolitic lava is erupted. But many granites are not visibly or even inferentially connected with surface flows, and therefore their magmatic origin is incapable of direct proof. It is with the origin of such granites that we are immediately concerned, and the nature of the evidence is examined in the following pages.

The first step towards building up a reasonable theory of the origin of the granitic rocks must be based upon the observation that they greatly predominate among plutonic rocks; but that among the volcanics the dominant type is basalt. Consideration of the distribution of these two contrasted types has led to the belief that granite can only originate where a granitic shell, the sial, is already in existence. This conclusion follows naturally from the fact that granites are restricted to the continental areas, but are absent from the ocean basins. In the latter only rocks of direct basaltic parentage can be formed. This seems to indicate that, in general, granites are not derived from basaltic magma by differentiation as has been advocated in the past, notably by Bowen. However, a particular variety of basalt termed tholeiite, of very wide distribution in continental regions, is distinguished by containing interstitial patches of residual glass. The latter may contain up to some 70 per cent of SiO_2 ; it is therefore thoroughly acid, and potentially granitic. If at the appropriate stage of crystallization it could be squeezed out of the basalt and allowed to crystallize independently, it would certainly form granite, microgranite or rhyolite according to the environment in which it cooled. This may

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well be the mode of origin of the occasional rhyolitic lavas intercalated among the dominant basaltic flows in many volcanic regions, and also of the minor granitic intrusions associated with the more abundant basic rocks. The quantity of granite originating in this way is small at best, and this mode of origin is quite inadequate to account for the enormous amounts involved, for example, in the orogenic belts in the western Americas.

It is probable that the formation of most granites has involved selective melting of the sial as advocated by Eskola, with the most soluble combinations of minerals melting first. The fact that quartz and alkali-feldspar (orthoclase and albite) in the approximate proportion of 1 : 1 : 1 combine to form the lowest-melting-temperature assemblage of rock-forming minerals is of paramount importance in this connection. Wholesale melting of rocks is termed **anatexis**; while the formation of new magma capable of intrusion at higher levels in the crust is termed **palingenesis**, implying rebirth of a new rock from the melting of a previously existing one. These processes can operate most effectively when the sial is thickened and crustal temperatures are raised in the root zones of mountain chains, thus accounting for the fact that plutonic rocks of the tonalite-granodiorite-granite association are largely confined to orogenic regions.

Petrologists seldom concern themselves very much with the probable sources of basaltic magmas, being content to assume derivation from the deeper crustal layers or from the underlying mantle. These are safely beyond the range of direct observation, at depths of the order of 20 or 30 km. In the case of granitic rocks, however, deep erosion in shield areas and in the cores of ancient mountain chains should expose or approach the levels where selective fusion of the granitic components might have occurred in the past. These are the great theatres of plutonic metamorphism, and in them even the granitic rocks are more metamorphic than igneous in character. This is so much the case, in fact, that the magmatic origin of "granites" in general has been questioned. A long and vigorous controversy has ensued between those upholding the magmatic origin of granites ("magmatists") and those regarding granites as essentially metamorphic products of granitization ("transformists"). There is no doubt that extreme and sometimes absurd claims have been made for both hypotheses; but now it is generally realized that the arguments of both schools of thought are valid if applied to granites of the right kinds and in the right environments. Probably no one has done more to bring about this realization than H. H. Read whose writings on the subject, collected in the book *The Granite Controversy*, form a critical commentary upon the whole problem.

ORIGIN OF THE GRANITIC ROCKS

Conclusions regarding the origin of any particular granitic rock must take into account the following:

(1) The structure of the granite mass and its relationship to surrounding rocks.

(2) Chemical composition of the granite, particularly by comparison with volcanic rocks of known origin.

(3) Mineralogical and textural features which provide evidence of the conditions of crystallization and cooling.

Each of these factors is examined in turn below, with particular reference to intrusive granites that may reasonably be regarded as essentially igneous rocks: the problems of granitization and the origin of "metamorphic" granites are mainly reserved for the final section of this chapter. Two points should be emphasized here. As this is a book on igneous petrology, the granites of metamorphic character are treated much more briefly than they deserve. Secondly, the division of granites into two categories is artificial and a matter of convenience only. All granitic rocks belong to a series with gradational properties.

(1) *Field relations and problems of granite emplacement*

In the following brief survey the main contrasts between granites at the two ends of the series are emphasized: those originating early in an orogenic cycle in deep levels of the crust are contrasted with granites formed later and generally at higher levels.

Very deep-seated granites are often concordant, with margins parallel to the foliation and/or bedding of the surrounding schists and gneisses. They display a wide diversity of form and size, from diffusely scattered augen measuring perhaps a centimetre or so in diameter, to thick sheet-like or lenticular masses of great lateral extent. Internal foliation is often conspicuous and there are clear indications either that the granites were in existence before folding and metamorphism occurred (*i.e.*, the granites were *pre-tectonic*); or that they came into being during folding, thus being *syn-tectonic*. The latter are more commonly discordant than the pre-tectonic granites and they may form irregular veins in "injection complexes," or be intimately mixed and interlaminated with metamorphic rocks of various kinds, forming the so-called **migmatites**. A uniformly high grade of regional metamorphism frequently extends through the granites and associated rocks of any particular area, indicating crystallization of all the rocks concerned under uniform conditions of temperature and pressure.

High-level granites in orogenic regions are generally much more obviously intrusive, forming plutons which are often of very large size. Contacts are either sharp or else there is a relatively narrow

PETROLOGY OF THE IGNEOUS ROCKS

transition zone between the granite and surrounding rocks. The latter often show little or no sign of regional metamorphism, though they may have been folded before intrusion of the granite. Many high-level granites are thus shown to be *post-tectonic*.

One of the most significant differences between deep-seated and high-level granites lies in the aureoles of thermal metamorphism which characteristically surround the latter, but not the former. The disharmony between the energy levels represented by such granites and the metamorphosed rocks into which they are intrusive is most significant. The thermal metamorphism which they have caused is but one of several phenomena which lead to the belief that high-level granites originate in general from the intrusion of hot and relatively fluid material. Whether this is regarded as magma is partly a question of definition and partly dependent upon facts of chemical composition which are discussed below. Granite is believed to be capable of diapiric¹ intrusion even when it is largely crystalline, the mechanism of intrusion being closely comparable with that of rock-salt which possesses the property of plastic flow. Unfortunately even a close study of flow-structures may fail to differentiate this condition from one involving dominantly fluid magma, because so much of the structural character of a granite is determined ultimately by late- or even post-crystallization movements. In view of the gradation that must exist and the impossibility of deciding the degree of fluidity in particular cases, it seems best to the writers to regard all granites in this category as magmatic in the wide sense.

The main problem arising concerns the means whereby the granite mass came to occupy its present position: and it is commonly argued that the **space problem** is less acute if replacement rather than displacement is invoked. Actually there is a space problem in both cases; and intrusion necessitating displacement of country-rock does not involve a greater space problem for magmatic rocks than for rock-salt or carbonatites. In Persia, Poland and elsewhere it is evident that plugs of rock-salt have punched their way through great thicknesses of overlying strata, producing little structural disturbance of the latter, and being driven upwards by their own gravitational potential only. The mechanism of emplacement may not be obvious from the structural evidence; but in the case of salt domes, at least there is no question as to their origin by displacement rather than replacement.

Space may be made for the emplacement of granite in several ways: probably in any specific case a combination of different processes was involved.

¹ Diapirism is literally a breaking through an envelope of country-rock. This may be achieved by intrusion of either fluid or mobile solid material, and gives rise to cross-cutting relationships.

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Harker long ago argued that magma could shoulder aside country-rock or lift the "roof" in making space for itself; mechanical displacement by cauldron subsidence or piston faulting has been already described; while piecemeal stoping is proved to have taken place in many cases where granite is seen to be charged with xenoliths.

The solution of the space problem presented by a particular granite depends upon detailed structural study of the rock mass. The nature of the problem is well illustrated by the Flamenville granite which outcrops on the coast of Brittany.

When first mapped it was thought that relics of original bedding could be traced through the granite as a *ghost stratigraphy* which linked up with the country-rocks on either side. If true, this would have made a most convincing argument for the granite having been formed *in situ* by replacement (*i.e.* granitization). Structural mapping has shown, however, that the country-rocks surrounding the granite have been intensely disturbed by forceful intrusion of the latter; while internal flow layers parallel to the margin suggest that the intrusion is mushroom- or pear-shaped with the widest part above a narrow stem.¹

The general characters of the main kinds of granitic intrusions have been indicated above in Part II, Chapter 2. Some idea of their variety may be gained from a study of the "granites" of Donegal, in Northern Ireland (Fig. 93). Four main types of intrusion are represented, each emplaced under different conditions. The earliest intrusion is essentially granodioritic and penetrated the country rock consisting of Dalradian metamorphic rocks in an intricate manner. The whole complex is migmatitic. Although the Dalradian rocks occur as inclusions enclosed within, and injected by the granodiorite, they have retained to a great extent their original positions in space. Preservation of such a "relict stratigraphy" indicates that the intruded rocks have suffered only slight displacement; but the fact that some rotation of xenolithic blocks and some relative displacement can be demonstrated in the field is regarded by Pitcher as evidence of the essentially magmatic character of the granodiorite.²

The Rosses granite complex³ cuts the older granodiorite. It has been formed by successive intrusions of four distinctive types of granite which are bounded by sharp and steep contacts giving polygonal outcrops. The boundaries may be regarded as ring-faults

¹ Martin, N. R., "The structure of the granite massif of Flamenville, Manche, north-west of France," *Quar. Journ. Geol. Soc.*, **108** (1953), pp. 311-42.

² Pitcher, W. S., "The migmatitic older granodiorite of Thorr District, Co. Donegal," *Quar. Journ. Geol. Soc.*, **108** (1953), pp. 413-46.

³ Pitcher, W. S., "The Rosses granitic ring-complex, County Donegal, Eire," *Proc. Geol. Assoc.*, **64** (1953), pp. 153-82.

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whose attitude has largely been controlled by master joints developed in each previously consolidated granite. The intrusion may be described as "permissive," *i.e.*, resulting from the inflow of magma into potential cavities induced by the ring-faulting. In this respect the Rosses complex is in complete contrast to the neighbouring

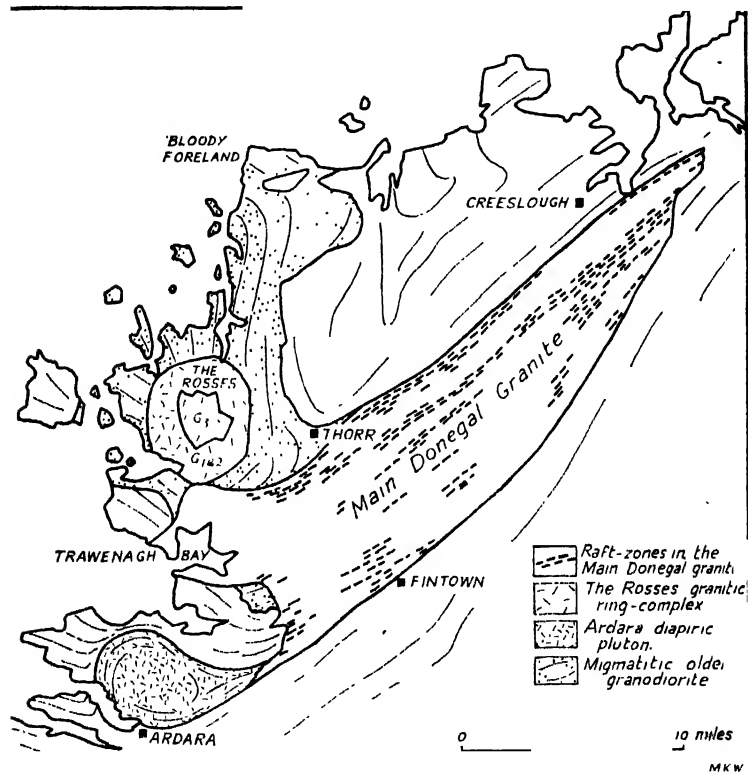


FIG. 93

Diagrammatic map of the Donegal Granite.

(Based on the work of Read, Pitcher and others.)

Ardara pluton¹ which is of the same general character as that of Flamenville, having been forcefully intruded as a diapir. The surrounding rocks have been pierced through and pushed aside in the process. It is interesting to note that this also is a multiple intrusion with a contaminated and more basic facies near the margin, crowded with xenoliths. This is a very common feature of

¹ Akaad, M. K., "The Ardara granitic diapir of County Donegal, Ireland," *Quar. Journ. Geol. Soc.*, 112 (1956), pp. 263-90.

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acid plutons and is displayed by many of the late-Caledonian intrusions of Scotland.

The Main Donegal Granite has the dimensions of a batholith. Much of the interest of this great mass is centred upon the trains of xenolithic inclusions which have been drawn out parallel to the length of the intrusion. The gradual divergence of the trains towards the south-west suggests that they were drawn out in this direction by lateral flow movements. Emplacement of the granite probably involved "lateral magmatic wedging";¹ but the process must have been immensely complex and long drawn out as evidenced by the structure and metamorphism of the surrounding rocks. These have been subjected to tight folding, and multiple metamorphism which in some respects has a "regional," rather than local "contact" appearance.

(2) *Factors involving chemical composition*

There can be no doubt as to the existence of magma of granitic composition: its reality is demonstrated—often with dramatic effect—every time rhyolitic lava pours out of a hole in the ground during a volcanic eruption.²

The similarities of composition of rhyolites, rhyodacites and dacites on the one hand, and of granites, adamellites and granodiorites on the other are so marked in some igneous provinces that they amount to virtual identity. Notable examples are provided by sodic rhyolites and riebeckite-granites in the province of the "younger granites" of northern Nigeria (see p. 193). Similarly the analysis of Arran pitchstone quoted on p. 224 corresponds almost exactly with that of a granite from the same area. These examples—and many others that could be quoted—come from volcanic regions where the magmatic origin of the intrusive rocks scarcely needs confirmation; it is only when the same chemical similarities can be shown to apply to genuinely plutonic "granites" that they form really valuable clues as to the magmatic origin of the latter. It is surely significant, therefore, that the chemical variation diagrams for O.R.S. lavas in Scotland apply almost equally well to the plutonic series of diorites, granodiorites and granites of the region;³ and also that lavas in southern California can be matched in composition with the various intrusions which make up the great batholith of that region.⁴

¹ Pitcher, W. S., and Read, H. H., "The Main Donegal Granite," *Quar. Journ. Geol. Soc.*, **114** (1959), pp. 259–305.

² See the very readable discussion by M. Walton, "The emplacement of granite," *Amer. Journ. Sci.*, **253** (1955), pp. 1–18.

³ Nockolds, S. R., "The Garabal Hill–Glen Fyne Complex," *Quar. Journ. Geol. Soc.*, **96** (1940), p. 451.

⁴ Larson, E. S., Jr., "Batholith and associated rocks of Corona, Elsinore, and San Luis Rey quadrangles, Southern California," *Geol. Soc. Amer. Mem.*, **29** (1948).

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This last example introduces another significant feature shown by many granitic rocks: within the limits of a single intrusion the rock may be remarkably homogeneous in texture and composition. Such homogeneity is to be expected of a magmatic rock, but is difficult to explain for one of this order of magnitude if a metasomatic origin is claimed.¹ These arguments apply to the whole range of granitic rocks defined in a wide sense; but there is one aspect of the composition of alkali-granite (*i.e.*, granite *sensu stricto*) which sets it apart from the rest. The proportions of quartz and alkali feldspars in many such granites vary within very narrow limits, and correspond closely to the low melting trough in the ternary system albite-orthoclase-silica (Fig. 72). This, of course, provides strong presumptive evidence that these particular granites originated from melts.

(3) Mineralogical evidence relating to temperatures of crystallization

It has been shown experimentally that the minimum temperatures of the melts in the Or-Ab-SiO₂ system can be drastically lowered under conditions of high water-vapour pressure, from about 950° C. for "dry" melts to about 650° C. with a pressure of 4,000 kg/cm².² By making certain assumptions regarding the thermal gradient (believed to be of the order of 30° C. for every kilometre increase in depth—but certainly variable from place to place in the crust), and also by assuming that the water-vapour pressure is equivalent to load-pressure, etc., it is possible to predict the levels in the crust where granite should become melted. In view of the number of variable factors involved it would be misleading to quote precise figures; but the experimental results indicate, with a high degree of probability, that melting must commonly occur at depths between about 10 and 20 km.

Very extensive research has been carried out by Tuttle, Bowen and others on the stability of the principal granite minerals, quartz and the alkali feldspars, in an attempt to use the evidence of the structural state of the minerals as indicators of the temperatures of their crystallization. Details relating to the inversion of α - to β -quartz have been described in the Mineralogy Section. Since there is a small but measurable difference between the inversion temperature for quartz that originally crystallized in the high (β) state compared with primary α -quartz, this difference should provide a

¹ See, *e.g.*, Chayes, F., "The finer grained calc-alkaline granites of New England," *Journ. Geol.*, 60 (1952), pp. 207-54.

² Tuttle, O. F., and Bowen, N. L., "Origin of granite in the light of experimental studies in the system NaAlSi₃O₈-KAlSi₃O₈-SiO₂-H₂O," *Geol. Soc. Amer. Mem.*, 74 (1958). This important memoir contains critical discussions on all of the experimental work on synthetic granites and the granite minerals carried out by Bowen and Tuttle and their colleagues.

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most valuable clue regarding the temperatures under which granites have been formed. In fact it has been found that the quartz in all the genuinely plutonic granites so far tested has the inversion characteristics of the primary low form. This might be taken to imply that *all* really deep-seated granites must have crystallized below about 600° C.¹ However, even the most ardent advocate of low-temperature granitization would scarcely claim this to be the case. It must be concluded, therefore, that during the long and involved history of cooling of any plutonic rock, quartz thoroughly *recrystallizes* and takes on the characteristics of the primary low variety regardless of its original state. This is a disappointing conclusion because the evidence fails just where it is most needed—for the deep-seated granitic rocks. Of course it provides added confirmation of the high temperature origin of many high-level and generally fine-grained granites, many of which contain quartz showing the primary β characteristics.²

Much the same situation arises in the case of the alkali feldspars, the stability relations of which have been described above. Briefly, a single phase of alkali feldspar (anorthoclase or sanidine as the case may be) is stable at high temperatures and occurs in acid lavas and rapidly cooled rocks of the appropriate composition. In a very restricted range of granitic rocks, for which the cooling rates have been somewhat slower, the place of sanidine may be taken by an unmixed perthite. As long as only one kind of alkali feldspar is present it is safe to assume that this must have crystallized originally at temperatures above the level of the solvus (see p. 168).

In most, if not all, plutonic alkali-granites, however, two separate and discrete kinds of feldspar occur, the one dominantly potassic (commonly microcline or microcline microperthite) and the other sodic. This may arise in one of three ways:

(1) Extreme unmixing, particularly if associated with recrystallization, could give rise to two alkali-feldspars from an original single phase precipitated at high temperatures. All evidence of unmixing would be lost in the subsequent recrystallization.

(2) Under conditions of very high water-vapour pressure, the granite liquidus can be depressed to a point where the solvus is intersected (see p. 168). It is theoretically possible for two alkali feldspars to crystallize directly from the melt under these conditions.

(3) The feldspars may have crystallized as sub-solvus phases under metamorphic conditions.

¹ The inversion temperature is $573 \pm 1^\circ$ C. at atmospheric pressure; but it rises to somewhat above 600° C. under very high pressures.

² See, e.g., Tuttle, O. F., and Keith, M. L., "The granite problem: Evidence from quartz and feldspar of a Tertiary granite," *Geol. Mag.*, **91** (1954), pp. 61–72.

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There is no certain way of distinguishing between the three possibilities, although the texture and environment of the granite may provide some indication. If, for example, the granite has a homogeneous composition, and uniform feldspar proportions are maintained throughout, then either of the first two explanations may apply. A heterogeneous distribution of feldspars, however, may well have resulted from granitization: further evidence of the latter may be provided by obvious replacement of K-feldspar by albite or *vice versa*.

(4) *Metamorphic granites and granitization*

Some of the more important characteristics of really deep-seated granites occurring typically in Pre-Cambrian shield areas have already been noted. Generally such granites display a harmonious relationship towards surrounding rocks both structurally and as regards metamorphic grade. This fact makes it difficult to assess the true nature of the granites concerned, and three possibilities have to be considered: firstly, the granites may be relics of an earlier generation of rocks of the same composition, *i.e.*, earlier granites, rhyolites, etc., which have been recrystallized during the current phase of regional metamorphism. The formation of granites by this simple process—recrystallization of material of appropriate composition—is illustrated by the Bréhat granite in Brittany,¹ which is believed to have resulted from recrystallization of a pile of eruptive rhyolitic rocks. Such recrystallization does not take place in the dry condition in all probability: it is much more likely to happen under the impetus of an advancing wave of granitization.

Secondly, the high grade metamorphism may have caused the segregation of granitic components from the surrounding rocks by processes relating to selective fusion. Thirdly, the granitic components may have been introduced into the rocks either as an accompaniment to the metamorphism or as a separate event. Whichever of these explanations applies, there is liable to be the most intimate mixing of the granitic and non-granitic rocks and minerals on all possible scales.

J. J. Sederholm² has described rocks of this kind as **migmatites**. The granitizing agents were termed by him "ichors," which possess properties intermediate between aqueous solutions and very much diluted magma, much of it being probably in a gaseous state. Detailed studies of such rocks, which are intermediate between metamorphic rocks and those of igneous aspect, have been carried

¹ Lafitte, P., *Bull. Serv. Carte Géol., France*, 53 (1955).

² Sederholm, J. J., "On migmatites and associated Pre-Cambrian rocks of South-Western Finland," *Bull. Com. Géol., Finlande*, No. 77 (1926), p. 89.

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out in a few areas in Britain, chiefly in the far north among the high-grade schists, gneisses and granulites of the Highlands. By tracing particular bands along the strike in the direction of increasing metamorphism, progressive granitization may be observed, involving the elimination of certain minerals including garnet and muscovite, while plagioclase crystals increase in size and become more sodic. The growth of large porphyroblastic feldspars is characteristic. The resulting rocks are appropriately termed **permeation gneisses**; they are metamorphic, in part metasomatic and represent a stage in the formation of rocks of granitic composition and of igneous aspect. Actual penetration of the permeation gneisses by granitic fluids followed, forming **injection gneisses**. In these, distinct interbanding occurs of quartzo-feldspathic layers of granitic composition, with biotite-rich layers representing original pelitic (argillaceous) material. These are typical migmatites. Not infrequently original sedimentary structures are faithfully preserved in the migmatites. With increasing metamorphism the distinction between layers of contrasted composition tends to become less pronounced, the rocks become "homogenized" by a process of metamorphic diffusion and in extreme cases no trace of the preceding stages outlined above have survived. There can be no doubt, in view of these facts, that migmatites and more advanced granitic gneisses are products of interaction between country-rock and "juices" (to be noncommittal) derived from granitic magma. The latter was considered by Sederholm and his adherents to be a necessary prerequisite for granitization. In other words, granitization can take place only if and where granitic magma is available.

There has been much speculation regarding the nature of the granitizing medium. According to the evidence of Tuttle and Bowen this should be in the form of a silicate melt provided there is a sufficient concentration of H_2O and the temperature is in the region of $600^{\circ}C.$ or more. There are many instances, however, where the evidence provided by the mineral associations in the accompanying metamorphic rocks suggests that granitization occurred at appreciably lower temperatures. In such cases relatively dilute aqueous solutions may be responsible (*cf.* the ichor of Sederholm). As an alternative it has been suggested that granitization may be brought about by the diffusion of ions through crystal lattices and particularly along crystal boundaries, this being in the "dry" state and without the intervention of a liquid phase.¹ The migration of ions is prompted by rising temperature and pressure combined: the activity of the ions depends upon the energy level. Detailed geochemical researches on contact phenomena have demonstrated that the various elements involved in the

¹ Backland, H., "The granitization problem," *Geol. Mag.*, **83** (1946), p. 110.

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metasomatism tend to migrate in a certain order. Where a particular element or combination of elements has accumulated at the limits of diffusion within the metasomatized rock, a zone distinctive in mineral composition results, and has been termed a "front." It has been shown that in some instances, at least, a **basic front**, involving Ca, Mg and Fe is first produced and moves in advance of the acid constituents.

The most enthusiastic exponent of the concept of metasomatic fronts in this country is D. L. Reynolds who has provided several detailed studies of contact phenomena shown by the Caledonian and Tertiary intrusions in Northern Ireland. Various references are given in the last part of this book. The "basic front" phenomena are well displayed in the dark zone rich in mafic minerals frequently developed at the contact between a basic xenolith and the magma in which it is enveloped. Indeed, basic and ultrabasic xenoliths of some observers are interpreted as being due to basic fronts advancing inwards during the course of the granitization of sedimentary rock.¹

The well-known porphyroblastic feldspars occurring in the contact zone surrounding granites are in the nature of a test case as regards granitization. A finely displayed contact between Malmesbury Shale and the granite at Sea Point just outside Cape Town provides critical evidence.² Here the shale has obviously been extensively migmatized, and impregnated by material of granite origin, the physical and chemical nature of which is not important for the moment. The darker, more shaly portions of the mixed rock are charged with large euhedral crystals of alkali feldspar, and all the evidence suggests that they have grown *in situ* within the modified country-rock as porphyroblasts formed by metasomatism. Feldspar porphyroblasts of identical appearance are not infrequent in fine-grained basic xenoliths enclosed in many granites such as Peterhead, Shap, and the Jersey granites, and they convey the impression that with a slightly higher degree of alteration of the matrix in which the porphyroblasts are enclosed, the resultant rock would be close to a porphyritic granite in at least appearance.

In consequence of this many geologists have come to regard the growth of feldspar porphyroblasts in xenoliths, and in country-rock neighbouring granites, as representing an advance guard of granitization. It is important to realize that the porphyroblasts are frequently indistinguishable from feldspar phenocryst in the associated granites: hypotheses designed to account for the former have quite naturally, therefore, been applied to the latter also. Reasoning along these

¹ See, e.g., Reynolds, D. L., "The Geochemical changes leading to Granitization," *Quar. Journ. Geol. Soc.*, **102** (1946), p. 389.

² Mathias, M., and Walker, F., "On a Granite-Slate Contact at Sea Point, Cape Town," *Quar. Journ. Geol. Soc.*, **102** (1946), p. 499.

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lines has led to the idea that the "phenocrysts" in some granites—even in granites of magmatic appearance—have arisen by granitization of essentially solid rock. It is scarcely necessary to point out that this is completely contrary to experience gained from a study of volcanic rocks in which phenocrysts represent early precipitates from the magma.

The explanation given above for the large feldspar crystals in apparently intrusive and magmatic granites is too inconsistent to be accepted without very serious reservations. We prefer to accept these large crystals at their face value, that is, as phenocrysts which may have had long and complex histories of crystallization. Thus, although they probably commenced growth at an early stage in the cooling of the magma, they doubtless suffered continuous modification by reaction with residual fluids. This would be consistent with the plutonic environment. Any residual magma would be somewhat enriched in volatile constituents and would permeate xenoliths and adjacent country-rock, at least for a limited distance from the granite. This fluid would, as it were, carry the conditions of physico-chemical equilibrium pertaining to the granite across the contact and into the country-rock. Provided that the composition of the latter were suitable, then similar mineral phases could grow in equilibrium with the magmatic residuum on either side of a granite/country-rock contact.

The significance of feldspathization in the contact aureoles of granites has been discussed in some detail because it illustrates how widely different interpretations can be put on even a minor (though critical) part of the evidence in the granitization controversy.

SYENITES AND RELATED ALKALI-RICH ROCKS

General introduction: matters of classification

THE rock-types containing little ("accessory") or no quartz at all include those formerly classified as Intermediate, Basic and Ultra-basic on the basis of arbitrarily selected ranges of SiO_2 percentages. Thus the Intermediate range extended from 55–66 per cent or thereabouts. The Intermediate rocks so defined were subdivided into three "clans" (R. A. Daly) according to the mineral composition, particularly in relation to the felsic constituents. Although we no longer use the SiO_2 percentage limitation we do regard the emphasis placed on felsic constituents as sound and fundamental. The lines followed are those already explained above and applied to the rocks containing *essential* quartz. The three clans corresponding to granites, adamellites and tonalites are:

Syenites	Syenodiorites	Diorites
with alkali-feldspars	(= Monzonites)	plagioclase
or feldspathoids	alkali-feldspars	dominant
dominant	and plagioclase	
	equal in status	

This three-fold subdivision runs through the whole gamut of igneous rock-types. The syenites and allied types are strongly *alkaline*; the diorites, etc., are, by contrast, equally strongly *calc-alkaline*. In the plagioclase-dominant series another factor has to be taken into account: variation in the *composition* of the plagioclase which theoretically ranges from oligoclase¹ to anorthite. This long range is arbitrarily divided (for reasons explained in due course) at An_{50} . Rocks containing dominant plagioclase more sodic than this constitute the Diorite Clan; those with more calcic plagioclase fall in the Gabbro Clan (see table below). The same essential distinction is carried across into the middle series which thus comprises Syenodiorites (with $\text{An}_{<50}$) and Syenogabbros (with $\text{An}_{>50}$).

¹ Albite is excluded as it is regarded as alkali (sodic) feldspar.

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	Alkali feldspars and/or feldspathoids	Plagioclase with alkali feldspars and/or feldspathoid	Plagioclase feldspars	
Over-saturated	C. Quartz-syenite	C. Quartz-syenodiorite (quartz-monzonite)	C. Quartz-diorite (tonalite)	An ₀
	M. Quartz-microsyenite	M. Quartz-micromonzonite	M. Microtonalite	An ₁₀
	F. Quartz-trachyte	F. Quartz-trachyandesite	F. Quartz-andesite	
Saturated	C. Syenite	C. Syenodiorite (monzonite)	C. Diorite	
		M. Micromonzonite	M. Microdiorite	
	M. Microsyenite	F. Trachyandesite	F. Andesite	
		C. Syenogabbro (kentallenite)	C. Gabbro Norite	An ₅₀
		M. Micro-kentallenite	M. Microgabbro (dolerite)	
Under-saturated	F. Trachyte	F. Trachybasalt	F. Basalt	
	C. Nepheline-syenite	C. { Theralite Teschenite		
	M. { Nepheline-microsyenite Pseudoleucite-porphry	M. { Microtheralite Microteschenite		
	F. { Leucitophyre Phonolite			
	Leucitic and nephelinic feldspar-free lavas	F. Leucite and nepheline-tephrite and basanite		An ₁₀₀

Table showing the chief rock-types in the Syenite, Syenodiorite-Syenogabbro and Diorite-Gabbro Series.

(C=coarse-grained; M=medium-grained; F= fine-grained. Note that "saturation" on the left-hand side of the diagram refers to the *felsic* constituents; saturation in regard to mafic constituents (olivine) is not taken into account. Anorthite-content of plagioclase is only relevant to the right-hand and middle columns.)

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The rocks in the alkali-rich series contain (ideally) no plagioclase, so such division does not apply; but instead a new complication arises through the incoming of the undersaturated equivalents of the alkali-feldspars—the feldspathoids—which may fill the role of alkali-feldspars to any degree, even to the complete exclusion of the latter. This kind of variation applies also to the Syenodiorite-Syenogabbro Series in which the alkali-feldspar (allowed by definition) may be “proxied” by nepheline or leucite to any extent.

Ideally the Diorite-Gabbro (or calc-alkali) Series contains no alkali-feldspar, therefore undersaturation in this sense does not arise—feldspathoids do not occur in these rocks, so that rock-types containing feldspathoids are restricted to the left-hand and middle columns of the appended table.

The table itself summarizes the facts stated above, and an attempt is made to include the medium- (M) and fine-grained (F) equivalents of the coarse-grained types (C) whose names (syenite, diorite, etc.) only have been used up to this point. Further details are, of course, included in the description of each clan which follows. One important line of variation which affects the rocks named in *all* three columns concerns the quantities of mafic constituents present. Variation in this respect is not indicated in the diagram, but is fully covered in the text.

In the following account syenites are described first, then the corresponding fine-grained types, usually occurring as lavas of the kinds named in the table. The basis of classification is the same for all three grain-size categories. There are three variables: (1) the degree of silica saturation, (2) the kinds of feldspar and feldspathoid, and (3) the amounts (rather than the kinds) of mafic constituents present. Ideally the feldspar is alkaline, but it is reasonable to allow some latitude for border-line cases, and accessory plagioclase may occur.

According to the degree of silica-saturation three groups may be distinguished:

- I. *Over-Saturated*. Quartz-syenites: rocks of syenitic aspect in hand specimens; syenitic composition confirmed by examination of thin sections, but containing quartz in an accessory capacity.
- II. *Saturated*, containing neither quartz nor feldspathoid. It is doubtful if any large rock-body is of this type: most syenites are either over- or under-saturated.
- III. *Under-Saturated*, differing fundamentally from the above by containing nepheline or other feldspathoid proxying for alkali-feldspar to any extent. The demarcation is clear and unequivocal: the occurrence of nepheline or leucite in these rocks marks the entry of a new phase in the mineral assemblage.

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The degree of under-saturation is indicated by the ratio of feldspar to feldspathoid, and three such degrees may be distinguished:

- (a) with alkali-feldspar in excess;
- (b) with feldspathoid in excess;
- (c) with feldspathoid exclusive—these rocks contain no feldspar.

Each of the categories in I, II and III above may be further subdivided on the basis of the kind of feldspar present: it may be

- (1) potassic (orthoclase or microcline);
- (2) sodipotassic (perthites); or
- (3) sodic (albite).

Finally, any one of the types so defined may vary in regard to the amount of mafic minerals which it contains. The sense in which the three varietal terms, repeated below, are used has already been explained:

- L — leucocratic;
- LM — mesocratic;
- M — melanocratic.

In the above summary statement the successive subdivisions have been deliberately "offset" on the page to suggest to the student that the examination of a given section of one of these rocks should be carried out systematically, when its position in the scheme of classification will be arrived at. Further, as an exercise, and in order to indicate the essentials, it is practicable to build up a symbol such as III(b) (2)M. Reference back to the foregoing indicates that the rock in question is an under-saturated syenite consisting essentially of nepheline in excess of perthitic feldspar, associated with mafic minerals in amounts above average for the type. Thus the symbol conveys *all* the essential facts of the mineral composition of the rock concerned. So, of course, does the selected rock-type name. A very simple calculation will show that, to cover the whole field of mineral variation, it might be considered necessary to recognize (and, incidentally, name) twenty-seven types of under-saturated syenites—a formidable number, especially when it is realized that the medium and fine-grained equivalents are still to come! As a fact, we quote these numbers as justification for doing everything possible to reduce the number of rock-names to a minimum.

Two important points concerning variation in mineral composition of syenitic rocks should be noted. Firstly, with regard to the nepheline-bearing types, the degree of under-saturation is widely variable: at the one extreme is the rock which contains only a small amount of nepheline of accessory status only, and therefore

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the rock is correctly termed "nepheline-bearing syenite": emphasis is laid on the essentially *syenitic* nature of the mineral assemblage. But at the other extreme occurs the rock in which the roles are reversed: nepheline is in large excess of alkali-feldspar, and in some varieties may be exclusive. The question arises as to whether it is appropriate to apply the general name "syenite" to a rock which lacks feldspar, one of the essential components of syenite as usually defined. Shand recognized this difficulty and advocated using the term "**syenoid**" (compounded of "syenite" and "feldspathoid") for rocks of syenitic aspect, but in which the place of alkali-feldspar is taken by a feldspathoid mineral. Syenoid is a useful term as it links the gradational members of the genuine syenites with the feldspar-free members of the ijolite series.

The second important line of variation concerns the ratio of felsic to mafic minerals. Most typical syenites are average in this respect; but some are extremely leucocratic, while others are notably mafic and are distinguished as **melasyenites**, or in the case of feldspar-free, under-saturated varieties, as **melasyenoids**. Actually this line of variation goes farther still, involving the gradual elimination of felsic constituents, the ultimate types being ultra-mafic and including some pyroxenites. These are right outside the syenite category of course; but they are mineralogically allied to syenites, they grade into them and occur in the same rock-bodies in some instances.

(I AND II. OVER-SATURATED AND SATURATED SYENITES

These are dealt with together for the reason stated above. Varieties containing (1) potassic feldspar, (2) sodipotassic feldspar, or (3) sodic feldspar may be distinguished.)

It is of historic interest perhaps to refer to the rock from Syene (Assuan) in Egypt, from which Pliny derived the word syenite. Actually the rock is a particularly handsome hornblende-*granite*. The rock selected as the type by A. G. Werner comes from Plauen near Dresden and is widely known as the **Dresden syenite**, and countless generations since Werner's day have been taught to regard it as representative of its kind. It is an over-saturated potassic syenite [type 1 (1)] containing some oligoclase—in some facies too much (up to 37 per cent of total feldspar) which places these varieties among the monzonites. In hand-specimens dull reddish, fluxionally arranged feldspars are obvious, the coloured silicate is identified as common hornblende on examining thin sections, while megascopic sphene, together with apatite and magnetite, are the usual accessories. In hand-specimens the feldspars are all uniformly brick-red, and this doubtless deceived early observers into thinking that only one kind of feldspar was present. Examination of thin sections shows that there are two kinds, orthoclase and oligoclase. Both are of the

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same crystal habit. It is rather easy, therefore, to over-estimate the amount of orthoclase in a section of Dresden syenite: the oligoclase is liable to be overlooked. It occurs as complex twins (Carlsbad-Albite) and by reason of the straight extinction of plagioclase of this composition the simple twinning is much more obvious than the lamellar Albite type.

Tyrrell¹ has described from Spitzbergen a syenite of the same general variety, but much nearer to the ideal. Quartz is present, but in insignificant amounts (3.3 per cent); microcline makes up 62 per cent of the rock, with oligoclase only 4 per cent; while biotite and augite in equal amounts make up the remaining quarter.

Equally typical syenites occur in alkali complexes in South, central and eastern Africa. One distinctive type contains a little interstitial green pyroxene, but the rest consists of platy, fluxionally arranged alkali-feldspars, dull red in the hand specimen, and giving the impression that every crystal is a Carlsbad twin.

(If albitites are really what the name implies, they conform ideally to an important syenitic type (II (3)) the saturated sodic syenites.) Albitite (H. W. Turner, 1896) was first recorded from Plumas Co., California; later records include Cape Willoughby, S. Australia, the Shetland Isles, Scotland,² and Kotaki, Japan. Albitite is virtually monomineralic. (It contains no free quartz although the SiO_2 percentage may be as high as 68. Over-saturated albitites (quartz-albitites) occur in dykes in the orogenic zone) in Japan, forming part of a dyke-phase associated with ultrabasic serpentinites, and including also albitites, trondjemites and other highly feldspathic rock-types. Albitite appears to be closely associated with serpentinites: "lenses" of the former occur embedded in the latter, but sheathed in jadeite. It will be remembered that jadeite is $\text{NaAlSi}_2\text{O}_6$, while albite is $\text{NaAlSi}_3\text{O}_8$, so that the jadeite shell probably represents desilicated albite.

As might be expected, syenites in which the feldspars are of less extreme composition are much commoner than those containing albite or orthoclase. The feldspars are sodi-potassic—occurring either as discrete crystal grains of albite and orthoclase or microcline, or intergrown as perthite or antiperthite. Examples are found in both the over-saturated and saturated categories. Over-saturated sodipotassic syenites are well represented by nordmarkite (W. C. Brögger, 1890), described originally from a locality near Oslo, Norway. The dominant component is micropertite accompanied by rather much quartz, which places some specimens of nordmarkite on or over the borderline between granites and syenites.

¹ Tyrrell, G. W., *Trans. Roy. Soc. Edin.*, 53 (1922), p. 225.

² Walker, F., "An albitite from Ve Skerries, Shetland Isles," *Min. Mag.*, 23 (1932), p. 239.

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Subordinate coloured silicates include aegirine, riebeckite or arfvedsonite, either alone or in company. If desired, therefore, distinction may be made between, say, aegirine-nordmarkite and riebeckite-nordmarkite. The nature of the coloured silicate(s) may best be indicated in this way. In Britain nordmarkite forms the upper part of the stratified laccolith forming Cnoc na Sroine in Assynt; it also occurs (with pulaskite) in the Ben Loyal Complex¹ and the Loch Ailsh Complex in Sutherlandshire. The modal analysis of the Ben Loyal nordmarkite is: 77 per cent feldspar (albite 47, orthoclase 30, therefore antiperthite) quartz 12, amphibole 9 with accessory sphene and magnetite.²

Perthosite (J. Phemister, 1926) described from a Scottish type-locality is a sodipotassic leucosyenite (type I (2) of the above classification) consisting almost wholly of perthitic feldspars and less than 5 per cent of mafic and accessory constituents. Even with 65 per cent silica the rock is quartz-free; it is exceptionally rich in alkalis, with Na₂O, 7.4, in excess of K₂O, 5.2 per cent.

Pulaskite is a very similar type, described from Pulaski Co., Arkansas, by J. F. Williams (1890); but much nearer home it forms the greater part of the Ben Loyal Complex in Sutherland, northern Scotland. This syenite also is markedly leucocratic, but with antiperthitic feldspars making up almost the whole of the rock, with a little aegirine and accessories.

A most distinctive type, both as regards megascopic and microscopic characters, is the very handsome Norwegian rock, widely used for ornamental purposes, and termed **larvikite** (or **laurvigite**) by Brögger (1890). The type is characterized by its coarse grain and distinctive feldspars, which, especially on polished surfaces, exhibit a beautiful blue schillerization. They were referred to as anorthoclase and were regarded as a cryptoperthitic intergrowth of orthoclase and oligoclase. In this respect they are akin to the feldspars in the lava-type, kelyte. Recently, however, it has been claimed that instead of a single feldspar-phase, there are *two* present—oligoclase and alkali-feldspar. The former is largely antiperthitic and it is exceedingly difficult to identify as it rarely shows twinning. Actually the plagioclase member is intricately twinned on the Albite Law, but on so fine a scale that the lamellae are invisible under ordinary conditions. To complicate the issue still further, the twinning imparts a monoclinic symmetry to the optical orientation of the crystals. This elusive internal structure finds outward expression in the development of a distinctive crystal habit under favourable conditions, well

¹ Read, H. H., in "Geology of Central Sutherlandshire," *Mem. Geol. Surv., Scotland* (1931), p. 174.

² Knorring, O. von, and Dearnley, R., "Note on Nordmarkite . . .," *Min. Mag.*, 32 (1960), p. 389.

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exemplified in the rhomb-porphyrries described below. Larvikite has always been regarded as a syenite, but in view of these facts, its retention in this family is questionable: some larvikites appear to be monzonitic in composition: the *substance* of oligoclase is present in the requisite proportion, whether it is intimately intergrown with the alkali-feldspar, or occurs in discrete crystals.¹ The specimen illustrated in Fig. 94 is *apparently* syenitic, however, and so far as superficial characters are concerned, it would be out of place among the monzonites. Larvikite grades into *lardalite*, in effect a nepheline-larvikite, but with albite in place of oligoclase.

On account of the dark colour of the strongly schillerized feldspars the mafic minerals are not conspicuous in hand-specimens, but they are exceptionally striking in thin sections. They form clots from which feldspars are excluded; the assemblage is not one which would be regarded as typically syenitic but would appear more "at home" in a gabbroic rock. The clots include schillerized titanite, iron-rich olivine, abundant apatites of large size and perfect shape, irregular masses of iron-ore from which crystals of lepidomelane tend to radiate. Late-formed finger-like outgrowths of pyroxene penetrate outwards into the surrounding feldspars. A second iron-ore occurs as a normal accessory in the form of octahedra of titanomagnetite embedded usually in titanite. It may well be that these extraordinary clots are of xenolithic origin.

Of the rock-types described above, albitite and perthosite are notably leucocratic, while nordmarkite and the "Dresden syenite" are mesocratic. Potassic melasyenites, as noted above, are rich in mafic minerals. For saturated potassic melasyenites the most appropriate existing name is *shonkinite* (Weed and Pirsson, 1895), from the well-known Shonkin Sag laccolith in the Bearpaw Mountains, Montana.²

The original shonkinite contains olivine; but for the purposes of classification it seems best to ignore this in the definition and to prefix the rock-name with "olivine-" when that mineral occurs: thus the type rock is olivine-shonkinite. Similarly some facies of the type rock contain a little nepheline. So long as the latter is merely an accessory, this variety may be appropriately termed nepheline-shonkinite; but if nepheline is of essential status, this name is no longer appropriate. The mineral assemblage orthoclase, nepheline, mafic minerals defines *malignite*, described below.

¹ Barth, T. F. W., "The Igneous Rock Complex of the Oslo Region"; Pt. ii, *Systematic Petrography of the Plutonic Rocks, Oslo* (1945), p. 76; but cf. Oftedahl, C., *ibid.*, Pt. ix, *The Feldspars* (1948), 71 pp.

² Barksdale, J. D., "The Shonkin Sag Laccolith," *Amer. Journ. Sci.*, **33** (1937), pp. 321-59; also Osborne, F. F., and Roberts, E. J., "Differentiation in the Shonkin Sag Laccolith, Montana," *Amer. Journ. Sci.*, **22** (1931), pp. 331-53.

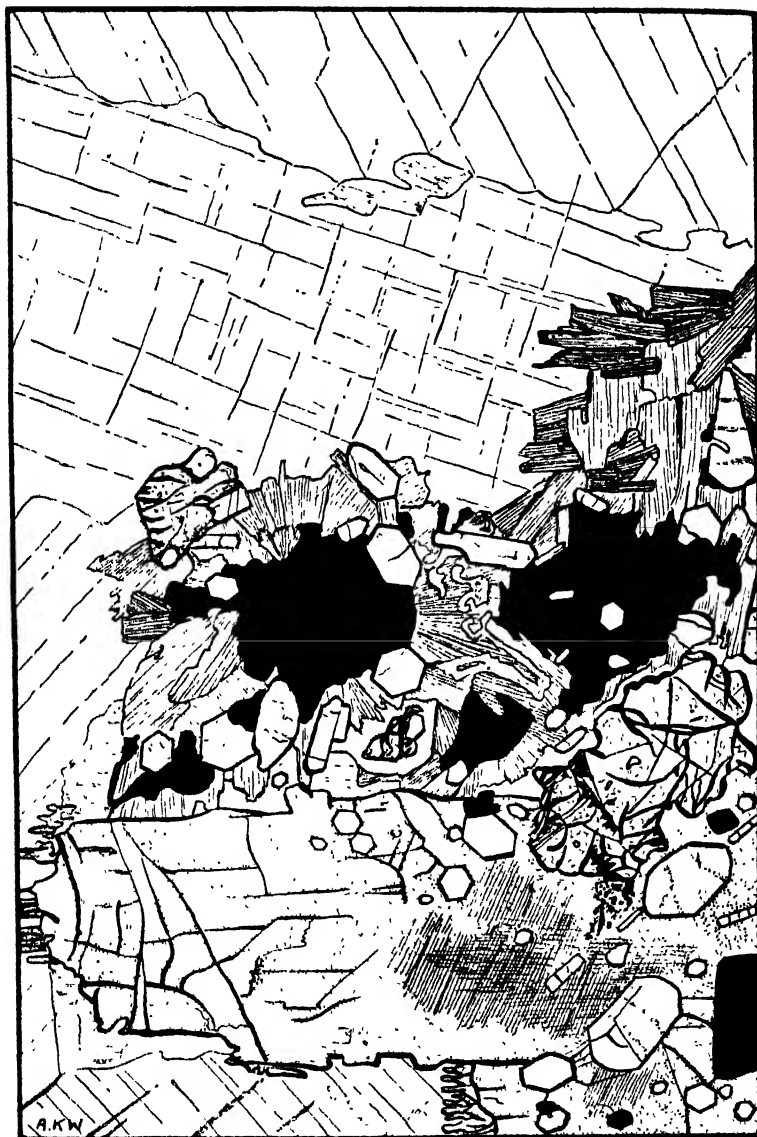


FIG. 94

Larvikite, Larvik, Norway.

Schillerized titanite, stippled; lepidomelane, lined; olivine, strong surface relief; titanomagnetite, black; apatite, plain. A little nepheline and much "anorthoclase" showing its cleavage.

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III. UNDER-SATURATED SYENITES (INCLUDING "SYENOIDS")

Within this important category mineral variation involves the kind(s) of feldspathoid present, the kind(s) of feldspar associated with the latter, the relative proportions of the one to the other, and the role of mafic constituents. The latter involves not only the amounts, but also the kinds of mafic minerals occurring. There are thus five variables, and ample scope, were we so inclined, for defining a large number of specific types. As a matter of interest there are actually more named types of nepheline-syenites than are found in any other category: we believe that many of these are superfluous in the sense that they differ only from better known rock-types in insignificant points of detail, which are adequately covered by appropriate qualifiers as indicated below.

Theoretically it would appear possible to distinguish two series of under-saturated syenites, one potassic, the other sodic; but actually this is impracticable. The significant minerals in this category are the feldspathoids, and therefore due weight is given to this fact by recognizing: (a) pseudoleucite-bearing types; and (b) nephelinic types. The former are exceedingly rare: the great majority of feldspathoidal syenites are nephelinic. The several mineral associations and the names of the corresponding rock-types are tabulated below:

- (a) with pseudoleucite accompanied by orthoclase: borolanite
- (a') with pseudoleucite only—"feldspar-free" types: fergusite,
missourite¹
- (b) with nepheline accompanied by potassic feldspar: malignite
- with nepheline accompanied by perthitic feldspars: foyaite
- with nepheline accompanied by K' feldspar and albite:
litchfieldite
- with nepheline accompanied by albite: mariupolite and
monmouthite
- (b') feldspar-free types with nepheline: the ijolite series; urtite,
ijolite and melteigite

In due course it will be shown that leucite-bearing lavas are much commoner than coarse-grained rocks of the same composition on account of the chemical instability of leucite under deep-seated conditions; and there is a closer and more complete similarity between the various nephelinic lavas and their leucitic analogues.

¹ There is a difficulty of terminology here, amounting almost to a contradiction in terms. Pseudoleucite consists largely of orthoclase; but these rocks are feldspar-free in the sense that they consist of pseudoleucite and mafic minerals only.

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The essential facts of the mineral composition of the main feldspathoidal syenites are indicated in the above table. Certain details of interest or significance are noted in the following brief account of individual types.

Types containing pseudoleucite

Borolanite is a most interesting rock although its status is uncertain. It is one of the few British type rocks and it is certainly distinctive enough to justify a specific name. The type area is the Loch Borolan Complex in Assynt, in the North-West Highlands of Scotland. It was originally described by Sir J. J. Teall who was impressed by the resemblance of certain prominent whitish crystals or aggregates to pseudoleucites.¹ Superficially the resemblance is close in some facies of the rock, but not in others. Thin sections show the whitish aggregates to consist of orthoclase associated with nepheline, usually decomposed. The abundant mafic constituents include melanite garnet, rich in titanium and associated with much sphene, pyroxene, greenish biotite, with purple fluorite as an unexpected accessory. The rock also contains polygonal areas said to be zeolitic aggregates after sodalite.

The felsic mineral association, orthoclase with nepheline, places borolanite among the nepheline-syenites, in this respect being close to malignite apart from the fact that it is rich in melanite. The systematic position of the rock depends, of course, upon whether it contains pseudoleucite or not. The uncertainty arises from the fact that borolanite occurs in a belt of country much affected by overthrust faulting; and Shand in a re-examination of the Loch Borolan Complex was led to believe that Teall's pseudoleucites were orthoclases which had been mechanically rounded during the earth-movements.²

With increasing melanite, borolanite grades into a type of melasyenite particularly rich in garnet, to which Shand originally gave the name ledmorite, from the nearby Ledmore River; but he subsequently withdrew the name in favour of melanite-malignite. (see the right-hand column of the above table and description of malignite below). The complex also contains in its basal parts a pyroxene-rich facies reminiscent of the corresponding ultramafic associate of melteigite described below.

There is no such uncertainty concerning **fergusite** (Pirsson, 1905) described originally from the Highwood Mountains, Montana. It is chemically similar to shonkinite but is under-saturated, with leucite (or pseudoleucite) taking the place of orthoclase. In the

¹ *Trans. Roy. Soc. Edin.*, 37 (1892), p. 163.

² *Ibid.*, "On borolanite and its associates in Assynt," 1909-1910, pp. 202-15 and 376-416.

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type rock pseudoleucite, consisting of the usual association of orthoclase with nepheline, makes up more than half the rock (65 per cent); aegirine-augite is the mafic constituent and magnetite the chief accessory. A stock of fergusonite has more recently been described from Tashkent in the U.S.S.R., while of greater interest are certain occurrences in the Roman volcanic province. Here fergusonite containing *fresh* leucite is represented by ejected blocks only. Only two minerals are present—leucite and pyroxene, with an unidentified whitish material. Evidently the rock is not of deep-seated origin and by reason of its mode of occurrence the expected alteration into pseudoleucite has not occurred. Chemically, as might be expected, outstanding features are low silica percentage (about 48 per cent) and high K_2O (9.5).

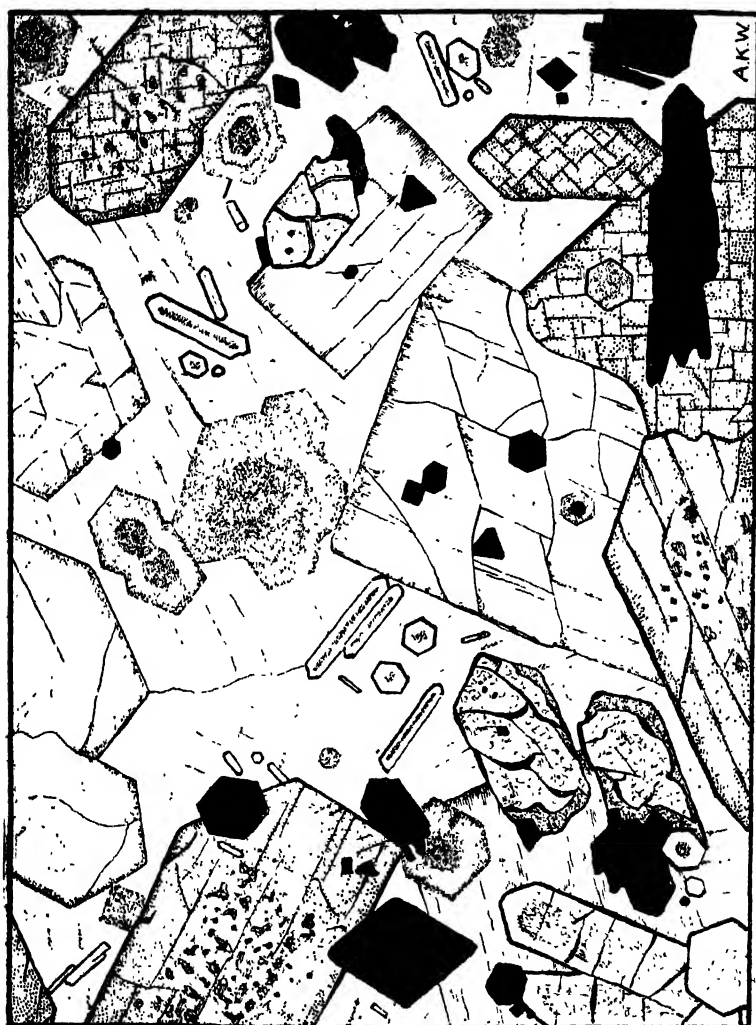
Missourite (Weed and Pirsson, 1896) also comes from the Highwood Mountains and is closely related to fergusonite, but is more melanocratic and contains olivine. As compared with fergusonite, missourite contains only 15 per cent pseudoleucite, about half the rock is augite, while olivine is also abundant. Thus missourite is olivine-melafergusite—the name missourite is not essential. Among the fine-grained leucitic lavas one type, olivine-leuamafite (olivine “leucitite”), is well known and is mineralogically equivalent to olivine-melafergusite.

Nephelinitic syenites and syenoids

Malignite (Lawson, 1896) takes its name from the type occurrence on the Maligne River in Ontario. The significant felsic mineral association is nepheline and orthoclase, in about equal amounts in the type rock. Clinopyroxene makes up about half the rock, while biotite, apatite and sphene also occur. With diminishing nepheline, malignite grades into shonkinite.

A malignite of particularly striking appearance is illustrated in Fig. 95. This is a well-known rock in teaching collections on account of the variety of minerals it contains. Two feldspathoids, fresh nepheline and euhedral crystals of a member of the hauyne-nosean group (nearer to nosean than to hauyne) together with interstitial poikilitic orthoclase are the felsic constituents. Euhedral zoned aegirine-augites with deeper coloured, more sodic outer zones accompany fewer and smaller olivines, which are partly serpentized. Prominent among the accessories are magnetite in unusually perfect octahedral crystals, and abundant apatites varying widely in size. Many of the larger apatites are distinctly cored. It will be realized that in squeezing this distinctive rock into the classification as a variety of malignite we are doing it less than justice. We know of no other rock in which the association of potassic feldspar (orthoclase) and sodic feldspathoid (nepheline) is so strikingly shown. The

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— 1 mm. —

FIG. 95

Olivine-malignite, Katzenbuckel, Odenwald. Aegirine-augite, stippled, olivine (three small crystals); nosean, turbid, showing zoning; nepheline, euhedral, clear; orthoclase, poikilitic, clear; apatite, several "cored"; magnetite, euhedral.

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rock is malignite to that extent: but it should be distinguished as an olivine-hauyne-malignite.

A malignite rich in melanite garnet was described from the banks of the Ledmore River in Assynt and was originally termed ledmorite by Shand; but the name was subsequently withdrawn in favour of melanite-malignite. Locally the content of melanite rises to 25 per cent. This rock is an associate of borolanite with which it has strong mineralogical ties.

Another very distinctive rock-type may be included in the malignite category, though Shand termed it a foyaite (see below): it contains microcline instead of orthoclase together with nepheline, mafic minerals and accessories. It forms a part of the Pilaansberg Complex in South Africa. The microcline is an apple-green variety resembling the well-known teaching material from Pike's Peak, Colorado. This feldspar contains barium in significant amounts.

III (2). Of the undersaturated syenites in which the feldspar is sodipotassic the type termed **foyaite** (Blum, 1861) is by far the most important, and is probably the most widely distributed of all the feldspathoidal syenitic rocks. Since the original discovery at Foya Peak, Serra de Monchique, Portugal, foyaite has been discovered in many alkali-rich complexes, notably in several in South Africa, described by Shand. These foyaite appear to conform to a type in which roughly one-quarter of the rock consists of nepheline, sodipotassic feldspar averages about 60 per cent, sodic pyroxene about 10, and sphene and other accessories 5 to 6 per cent. The alkali feldspar forms flat tabular white crystals contrasting with the red-weathering nepheline and the black lustrous prisms of aegirine. The habit of the nepheline varies: in some cases it is interstitial to the feldspars, in others it is euhedral, forming the characteristic stumpy hexagonal prisms, the identification of which, in hand-specimens, is easy by reason of the quarter-inch square and six-sided sections visible on the weathered surface. Occasionally significant accessories occur in a foyaitic mineral assemblage. Thus in the Pilaansberg ring-complex in the Transvaal Shand records a white sodalite-foyaite which "glitters in the sunlight like marble." This is transitional into sodalite-syenite, considered below.

Litchfieldite (Bayley, 1892) resembles a leucofoyaite in bulk composition, but differs in that it is a *two-feldspar* instead of a one-feldspar nepheline-syenite. Albite is dominant, but is accompanied by microcline. In the mass the rock is whitish, with scattered black lepidomelane crystals and lemon-yellow patches of cancrinite, while sodalite is also present. The type-rock is a cancrinite-sodalite leuco-litchfieldite.

III (3). The most strongly sodic syenites consist of nepheline associated with albite and mafic minerals; and two named varieties

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may be noted. **Mariupolite** (Morozewicz, 1902) contains albite, about three-quarters of the rock, nepheline 13 per cent, aegirine 7.5 and lepidomelane 4 per cent. The important feature is the minor role of nepheline. We consider that mariupolite is sufficiently well defined in terms of the feldspar-feldspathoid relationship only: the type-rock contains aegirine as the chief mafic mineral; but there is room in the scheme for, say, riebeckite-mariupolite as well as for aegirine-mariupolite, and this may be regarded as a kind of test case. Mariupolite *could* be defined by the mineral assemblage albite (in large excess), nepheline *and* aegirine. If this is done a new name must of necessity be found to cover those mariupolite-like rocks which contain a mafic mineral other than aegirine. The necessity for introducing new names does not arise if it is agreed that mariupolite is defined by the feldspar-feldspathoid relationship *alone*, without specifying the mafic minerals which may be present in different varieties: they are most satisfactorily covered by prefixing the name (or names) of the mafic constituents to the general rock-name mariupolite, as above.

By contrast, the second essential assemblage involves the same two minerals but with the roles reversed: nepheline predominant, albite accessory. This assemblage is represented by **monmouthite** (Adams, 1904), named from Monmouth Co., Ontario. Nepheline makes up more than three-quarters of the rock. In the type rock an amphibole is the mafic constituent; but aegirine-monmouthite has also been described.

The mineral which most commonly acts as proxy for nepheline is the allied silicate, sodalite. As noted above, sodalite may occur, in a very subordinate capacity, in such types as foyaite. As soon as it attains to the status of an essential component, a new name is required. **Ditróite** (Zirkel, 1866) from Ditró in Transylvania, is such a rock. In it nepheline may be as abundant as feldspar (microcline-microperthite in the type-rock), and is accompanied by both sodalite and cancrinite. The latter is apparently secondary after nepheline. While a green biotite occurs in some specimens from the type-area, aegirine-augite rimmed with very dark green arfvedsonite is more characteristic. Ditróite is a sodalite-bearing nepheline-syenite. Obviously with the roles of the two feldspathoids reversed, the rock would be nepheline-bearing sodalite-syenite. True **sodalite-syenite**, containing no nepheline, lies at the end of this line of variation, and is represented by a rock to which this name was originally given, occurring at Julianehaab in Greenland. Sodalite is not particularly abundant in it, amounting to some 8 per cent, but is the only feldspathoid present. Much of it is interstitial to lath-shaped sections of microcline-microperthite, but some of it

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occurs in the form of euhedral rhombdodecahedrons embedded in the feldspar. It is thus obviously primary. Similar sodalite-syenites have been described from Luishe in the (late) Belgian Congo, and from Nyasaland. Sodalite-rich syenites are rare; but an example from St. Hilaire Mountains, Quebec, contains up to 70 per cent; while a facies of the sodalite-syenite in Greenland grades into an almost pure sodalite rock, of an intense blue colour.¹

It is appropriate to consider **analcite-syenite** next, if for no other reason than the close similarity between sodalite and analcite. Both are cubic, colourless and have low refractive indices, and there is no easy optical method of differentiating between them: a micro-chemical test is the only certain means of distinction. "Analcite-syenite" used without qualifiers implies the association of analcite with alkali-feldspars and coloured minerals. G. W. Tyrrell² has described a coarse-grained analcite-syenite from an Ayrshire locality. The dominant mineral is anorthoclase rimmed with orthoclase, but it is accompanied by some plagioclase. The dark minerals are titan-augite rimmed with aegirine, and barkevikite. All these minerals are embedded in an abundant matrix of analcite. In this locality the analcite-syenite forms the upper part of a composite sill, the lower part of which is a basic analcite-gabbro (crinanite). This suggests that analcite-syenite is a differentiate of a basic magma carrying the constituents of analcite. The occurrence of thin veins of the former in a differentiated teschenite sill at Hallcraig in Fifeshire³ supports this hypothesis, as teschenite is closely similar to crinanite.

Nepheline-syenite may contain cancrinite. In many instances its relations to the nepheline indicate that it is an alteration product of the latter; but in other cases it is not only more abundant, but it has the status of an essential component, and in appearance seems to be primary. By analogy with other types considered above, such rocks are cancrinite nepheline-syenites or **cancrinite-syenites**, depending upon the amount of cancrinite in the rock.

Nepheline-syenites and more particularly the pegmatitic facies are noteworthy on account of the wide variety of accessories they contain. Many of these are titanium and zirconium silicates. One of the more striking is eudialyte, which in some rocks is obvious in the hand-specimen, but which rarely ranks as an essential component. **Eudialyte-nepheline-syenite**, named chibinite (Ramsay, 1894) from a locality in the Kola Peninsula in the U.S.S.R. is a very striking-looking rock in the hand-specimen: it consists of black aegirine

¹ Rosenbusch, *Mikroskopische Physiographie* (1907), p. 204.

² *Quar. Journ. Geol. Soc.*, **84** (1928), p. 540.

³ Campbell, R., "The Braefoot Outer Sill, Fife," *Trans. Geol. Soc. Edin.*, **13** (1933), p. 148.

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prisms and bright cherry-red eudialyte¹ in a matrix of nepheline and microperthite.

Finally, reference may be made to another rare type—**corundum-syenite**—which occurs in the Haliburton-Bancroft area in Ontario² and in the Ural Mountains. The rock from the latter locality contains no nepheline, but it falls in the category of under-saturated syenites in virtue of the presence of corundum. This mineral makes up nearly a quarter of the rock, most of the rest being microperthite, the third component being biotite in very small amount. The Canadian specimens on the other hand are corundum-bearing nepheline-syenites, often of pegmatitic facies. The origin of the corundum-bearing syenites is discussed later.

III (c). *Nephelinic types containing no feldspar* are completely under-saturated so far as felsic constituents are concerned. The only variability involves the kinds and proportions of mafic minerals. Three rock-types of wide distribution and closely associated in the field fall into this category. Their essential relationships are shown in the appended table. As the constituent minerals are essentially the same in all three, and only the proportions vary, it is obviously possible to use only one rock-name, that of the central, "average," type, ijolite, distinguishing the melanocratic variety as **mela-ijolite**, and the leucocratic variety as **leuco-ijolite**. Pulfrey has suggested dividing the series rigidly at the points in the colour-index scale as shown by the figures on the right.

	<i>Colour Index</i>
nepheline-rock	
_____	10
urtite—leuco-ijolite	
_____	30
ijolite—ijolite	
_____	70
melteigite—mela-ijolite	
_____	90
pyroxenite	

Urtite (Ramsay, 1894), a whitish rock of syenitic aspect, is named from the type-locality Lujaur-Urt in the Kola Peninsula, U.S.S.R. Nepheline makes up 85 per cent of the rock, the rest consisting of mafic minerals, in most instances (though not exclusively) aegirine. Although the colour index is so low (only 15), the silica percentage

¹ This is a very different colour from the red altered nepheline seen in many foyaites: it is a "raw" deep pink. Further, the eudialyte has a vitreous lustre: the altered nepheline is dull.

² Adams, F. D., and Barlow, A. E., "Geology of the Haliburton-Bancroft area, Ontario," *Geol. Surv. Mem. Canada*, No. 6 (1910).

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FIG. 96

Analcite-Ijolite, Neudorf, Schreckenstein, Austria. The chief components are titanite and nepheline, both in well-formed crystals, especially the latter. Analcite is interstitial to these, and is partly altered to a zeolitic aggregate. Apatite and iron-ore are prominent accessories. The nephelines are traversed by canal-like fractures, while the cleavage is accentuated by incipient alteration.

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is very low (45 per cent), while, as might be expected, both alumina (30 per cent) and alkalis (20 per cent), chiefly soda, are both exceptionally high. In the type rock the mafic mineral is again aegirine; but as defined above, it might be any of the soda-iron amphiboles or pyroxenes.

A small enough specimen of urtite may be pure nepheline rock, and in the mass the name which should be applied to a rock of this composition is nephelinite—there is no other available. Unhappily this name is widely used for the nephelinic lavas referred to below under “ijolite.”

Ijolite (Ramsay and Burghell, 1891) originally named after a locality in Finland consists of nepheline making up approximately half the rock, the rest consisting of unspecified coloured silicates and accessories. The type-rock contains aegirine; but an ijolite described from Songo, Sierra Leone, contains a pyroxene within the diopside-hedenbergite range.¹ Ijolite is much more widely distributed than most of the types so far noted: it occurs in association with nepheline-syenites (usually foyaites) in many alkali complexes, for example that at Spitzkop, Sckukuniland, where it is represented by a handsome coarse-grained rock in which black prismatic aegirines are embedded in a matrix of dull red nepheline with a little bright yellow cancrinite. It may be noted that certain important very basic feldspathoidal lavas consist of the ijolitic mineral assemblage: there can be little doubt that they are comagmatic. With decreasing mafic minerals ijolite grades into urtite, and in the opposite sense into **melteigite** (Brögger, 1921). In the type-rock nepheline makes up about a quarter of the whole rock, the rest being mafics and accessories. Pyroxene (aegirine or aegirine-augite) is strongly dominant: the rock is exceedingly melanocratic and is at the opposite end of the scale to urtite, the two being complementary.

MICROSyenITES

Microsyenites are the medium-grained equivalents of syenites on the one hand and trachytes, leucitophyres and phonolites on the other. They are not common rocks, and in this respect are less important than the other grain-size groups. Several alternative names have been applied to them in the past, some of general significance, others with more specific meaning. As most of these rocks are porphyritic, with more or less prominent phenocrysts of orthoclase, they have been collectively grouped as “porphyries”

¹ Baker, C. O., Marmo, V., and Wells, M. K., “The ijolites at Songo, Sierra Leone,” *Col. Geol. and Min. Resources*, 6 (1956), pp. 407–15.

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(as distinct from "quartz-porphyries," which are Acid rocks of the same status, but of granitic composition, of course). In a former edition of this book we used "microsyenite" for aphyric types, and "syenite-porphyry" for those with phenocrysts. It seems better to call all these rocks, regardless of texture, by one name, and, by analogy with the granitic rocks, **microsyenite** seems the obvious choice. The two textural variants are then "aphyric microsyenites" and "porphyritic microsyenites," the meanings of which are self-evident.

As regards mode of occurrence, these rocks are intrusive; but the mere fact that a syenitic rock occurs as a sill or dyke does not automatically entitle it to the name microsyenite. The latter should only be used if the rock is of medium grain; if of fine grain, it must be termed trachyte, if preferred, with the prefix "intrusive" to remove any possible doubt.

Certain of these rocks are definitely potassic, and differ from the corresponding syenitic and trachytic types only in grain size. **Potassic microsyenites** often have a distinctive texture referred to as orthophyric, in effect a coarser trachytic texture resulting from the close packing of short stout prisms of orthoclase. Orthoclase is the dominant mineral, often occurring in two generations, and accompanied by hornblende, biotite, and, in the more acid examples, by a little interstitial quartz. A particularly good example of a porphyritic potassic microsyenite occurs at Goodsprings, Nevada, and is noteworthy for the beautiful phenocrysts of reddish orthoclase, usually of simple form, for which it is famous. The perfect crystals from this rock are to be found in all good teaching collections. The groundmass also consists chiefly of orthoclase, and the amount of coloured minerals is small, as in most syenitic rocks.

If the magma from which these potassic microsyenites were formed became desilicated, leucite would crystallize at high temperature, but under intrusive conditions it would invert into pseudoleucite with falling temperature, and the rock would contain pseudoleucite and orthoclase (the proportions depending upon the degree of under-saturation achieved), accompanied by coloured silicates of the appropriate composition. Such a rock would be exactly analogous with pseudoleucite-syenite on the one hand and leucitophyre on the other and forms the connecting link between these two types. It is equally rare, but examples occur in the alkali-complex at Magnet Cove, Arkansas, and others have recently been noted, under the names "leucite-porphyry" and "pseudoleucite-porphyry," associated with "orthophyres" and potassic trachytes in an Eocene volcanic area in the Pambak Mountains, Armenia.

Sodic and sodipotassic microsyenites compare closely with corresponding coarse-grained rocks, with which they are intimately

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associated in the field, either as marginal phases or as dykes and sills. Provided the general principle of naming such relatively finer-grained facies of distinctive plutonites after the latter as explained above (p. 247) is accepted, a clearer picture is evoked by using such specific names as microlarvikite, microfoyaite and porphyritic micropulaskite, for example, and further description is scarcely necessary. Unfortunately even today names are applied to such rocks which give no indication whatever of their relationship to other named parental types. Thus in a recent revision of the classic Oslo petrographic province, the relatively finer grained larvikites are called akcrites.

The Oslo district is also famous for its "rhomb-porphyrries,"¹ so named by von Buch (1870) on account of the distinctive shape of the cross-sections of the feldspar phenocrysts. These rocks occur as lava-flows and dykes, and, although some are definitely fine-grained, most measured specimens lie just over the boundary between fine and medium, and are therefore appropriately considered here. Further, the composition of the rocks varies widely: some are monzonitic; but others, with a very low content of An (less than 1 per cent in extreme cases), are clearly microsyenitic. The phenocrysts are antiperthitic, the dominant partner being oligoclase or andesine, rimmed with and veined by "anorthoclase" to the extent of 30 to 50 per cent of the whole. The groundmass feldspars are microlites of anorthoclase, approximately $Ab_{52}Or_{45}An_3$. In mineral composition, therefore, rhomb-porphyrries must closely resemble larvikites, with which they are presumably consanguineous. We consider "porphyritic microlarvikite" to be an appropriate name for them.²

Under-saturated types of microsyenites are well exemplified by the microfoyaite type, which is the mineralogical equivalent of foyaite, in the medium grain-size group. The essential components are therefore alkali-feldspar and nepheline associated with such minerals as aegirine, aegirine-augite or riebeckite. These mineral names may be prefixed to the rock name to give added precision. Riebeckite-microfoyaite grades into the commonest type of phonolite as the grain becomes finer; aegirine-microfoyaite similarly grades into typical tinguaitite. Shand has employed "microfoyaite" to distinguish certain fine-grained facies of foyaite occurring in the alkali-complexes of Spitzkop and Pilaansberg in South Africa. This precedent may well be followed without rigid adherence to a fixed limit

¹ Oftedahl, C., "Studies on the Igneous Rock Complex of the Oslo Region. vi. On Akerites, . . . and Rhomb-Porphyrries," *Skr. utg. av Det Norske Vidensk.-Akad. Oslo.*, No. 1 (1946), p. 37.

² Specimens of several types of rhomb-porphyrries were brought to this country by the North Sea ice and are spread along the Yorkshire coast as erratics familiar to glaciologists.

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FIG. 97

Microcline, Spitzkop, Sekukuniland, South Africa. Aegirine, stippled; nepheline clear, but zoned occasionally; apatite, with "varnish" of cancrinite (top, left); calcite with reaction rim of cancrinite (bottom, right).

of grain size: the thing that matters is that there is an appreciable difference in texture between the parent rock and the facies which it is desired to indicate as something different. A case in point is illustrated in Fig. 97.

SYENITE APLITES AND PEGMATITES

Syenites are rare rocks, of very limited distribution. It follows therefore that syenitic rocks analogous with the granite-pegmatites

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already considered, will be very uncommon. They do occur, however, particularly in S. Norway where they are important on account of the rare minerals some of them contain.

In regard to syenite-aplites, bearing in mind the nature of the syenites, and the improbability of there being extensive bodies of syenitic magma, it follows that the alkali-rich residua, comparable with granitic aplites, must be exceedingly rare. Further, as syenites themselves are notably leucocratic rocks, it must prove very difficult to be certain that a given rock, suspected of being a syenite-aplite, is really such, and not merely a fine-grained facies of a normal syenite. The **bostonites**, named by Rosenbusch (1882) from Boston, Mass., are fine-grained rocks, occurring as dykes, and closely resembling trachytes in appearance. Mineralogically and texturally these rocks are leucocratic microsyenites or intrusive trachytes as the case may be. If the relations of these dyke-rocks to the parent mass are the same as those obtaining between granite-aplite and granite, then they are syenite-aplites; and if they match the type bostonites in the details of their mineral contents and texture, then they are syenite-aplites of the Boston type, or bostonites.

THE TRACHYTES AND RELATED TYPES

The fine-grained members of the alkali series of rocks of Intermediate composition are the trachytes, which thus correspond in mineral content with the syenites. The name trachyte was first applied to all volcanic rocks that, on account of vesicularity, were rough to the touch. Later the term was restricted, first to lavas of Intermediate composition, and later to those containing dominant alkali-feldspar. Trachytes are readily divided into three groups on the silica-saturation principle: (a) those containing free silica—the quartz-trachytes; (b) those which are exactly saturated, containing neither quartz nor feldspathoid—the ortho-trachytes; and (c) those which contain an unsaturated mineral of felsic type, including nepheline, leucite and other feldspathoids.

(a) *Quartz-trachytes and Orthotrachytes*

Repetition is prevented by describing these two categories together: the sole difference is the occurrence of accessory quartz in the former. In theory, the same degree of latitude is allowed as for the syenites: quartz may occur up to 10 per cent of the whole rock. It never occurs as phenocrysts, but is restricted to the groundmass, where it may occur as interstitial grains lying between the feldspar microlites, or it may form micropoikilitic patches, optically continuous over an area which may include parts of a number of feldspar microlites: in other words, the quartz forms little "lakes" in

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the slide into which the ends of a number of microlites of feldspar may penetrate. Thus in so far as the mere names of the component minerals are concerned, there is nothing to choose between quartz-trachyte and rhyolite. Actually the amount of quartz in the former is much smaller than in an average specimen of the latter, while texturally they are very different. Although a little glass may be present, trachytes are not notably vitreous. Most trachytes are porphyritic, with phenocrysts of feldspar, often of large size. The groundmass is typically microlitic: *i.e.*, it consists of closely packed microlites, lath-like in shape, and several times longer than they are broad. On account of flow movements in the magma they are often in parallel alignment. At its best, therefore, the *trachytic texture* involves a close packing of feldspar microlites which often exhibit flow structure (Fig. 98). The amount of coloured silicates in trachytes is small: they are therefore light, both in colour and weight.

In some instances it is possible to subdivide further, on the basis of the dominant alkali, and to classify the rock as a potassic or a sodic trachyte respectively. In the former, the feldspar is sanidine, ideally, and is accompanied by ordinary biotite, clino-pyroxene or hornblende. In the latter the feldspar may be albite, but is commonly a sodi-potassic type, while the coloured silicates are the distinctively coloured soda-iron amphiboles and pyroxenes, already noted in the syenites of comparable type: *viz.*, aegirine, riebeckite, cossyrite, etc. It is important, however, to note that in some trachytes, while the feldspar is dominantly potassic, the coloured minerals may be strongly sodic. One such rock has been described, consisting of nearly 90 per cent of sanidine, the rest consisting of aegirine and accessories. It is customary to prefix the rock-name with that of the coloured mineral, and to speak of biotite-trachyte, hornblende-trachyte, augite-trachyte, riebeckite-trachyte, or aegirine-trachyte.

The Drachenfels trachyte enjoys the reputation, among trachytes, of the Dresden syenite among the syenites; but like the latter it is not an ideal type—it contains a good deal more oligoclase than a typical trachyte should. It is noteworthy for the extremely large size of sanidine phenocrysts, which measure over an inch in diameter. Much smaller plagioclases also occur, embedded in a light cream-coloured matrix.

A more typical trachyte is illustrated in Fig. 98 which represents a rock of this type from the Solfatara volcano near Naples.

Trachytes occur in the closest association with basalts, not only in recent lava fields, but also in those of earlier geological periods. Thus among the Ordovician volcanic rocks in North and South Wales, and those of Devonian and Carboniferous age in Devon and Cornwall, trachytic rocks occur, occasionally as lava flows, but more commonly as minor intrusions. As a rule in these ancient eruptives



FIG. 98

Trachyte, Solfatara, Naples. The principal phenocrysts are of sanidine, often forming large crystals. Oligoclase also occurs in less abundance. The groundmass is almost wholly composed of sanidine laths, though sparse interstitial patches of isotropic glass occur. Coloured minerals include pale brown hornblende, and greenish augite, sometimes in poikilitic plates (not shown).

the coloured minerals are so much altered that it is impossible to state their original nature. They are often somewhat over-saturated, and while in many the feldspar is pure albite, in some it is a strongly sodi-potassic type. In this country it is customary to refer to these albitic trachytes as **keratophyres**, particularly when the associated basic lavas are also albitic—the so-called spilites.

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Equally strongly sodic trachytes, probably of Carboniferous age, occur in the Eildon Hills near Melrose, southern Scotland.¹ These, however, are in a much better state of preservation, and contain the sodic amphibole, riebeckite, in the characteristic "mossy" aggregates. From the Tertiary volcano of Mull,² a similar trachyte occurs, but contains approximately equal amounts of aegirine-augite and riebeckite.

A wide variety of trachytic rocks have been described from the neighbourhood of Mount Kenya, Kenya Colony.³ In the direction of increasing acidity they grade into sodic rhyolites, and in the other, with the incoming of nepheline, into phonolites. They closely resemble the suite described by H. S. Washington from Pantelleria, near Sicily,⁴ particularly in containing the rare sodic amphibole, cossyrite. Those rocks to which the name pantellerite was originally applied are apparently thoroughly acid (with silica percentage up to 70), and are distinctive by virtue of the occurrence of phenocrysts of anorthoclase and cossyrite, usually associated with pyroxenes. Some contain quartz phenocrysts, and are evidently sodic rhyolites, of special type; but in others quartz is no more than accessory, and the groundmass is trachytic. Such rocks, both in the type-locality and in Kenya, are aptly called **pantelleritic trachytes**. As a point of detail it may be noted that the quartz-free members of this Central African suite contain small amounts of Fe''-rich olivine. This is also true of the trachytes which occur in the puy of Auvergne: S. J. Shand⁵ has shown that the well-known rocks from Mount Doré contain about 2 per cent of this mineral, which is a rare constituent of coarse-grained rocks of like composition.

White-weathering trachytes contrast spectacularly with the finely exposed modern basalts of the oceanic islands such as Ascension in the South Atlantic. There is little doubt that in Carboniferous times the Midland Valley of Scotland must have presented much the same aspect. In what has survived of this lava-field trachytes form occasional flows, but more commonly occur as minor intrusions and vent infillings, due doubtless to the high viscosity of the magma. This is true not only of this lava-field, but also of those elsewhere and of other ages. Further, many of these rocks exhibit some degree, usually slight, of under-saturation, and are therefore considered below.

¹ Lady M'Robert, *Quar. Journ. Geol. Soc.*, **70** (1914), p. 303.

² "Tertiary and Post-Tertiary Geology of Mull," *Mem. Geol. Surv.*, Scotland, (1924), p. 191.

³ Campbell Smith, W., "A Classification of some Rhyolites, Trachytes and Phonolites from part of Kenya Colony," *Quar. Journ. Geol. Soc.*, **88** (1931), p. 212.

⁴ "The Volcano and Rocks of Pantelleria," *Journ. Geol.*, **21** (1913), p. 683; and **22** (1914), p. 16.

⁵ *Geol. Mag.* (1935), p. 86.

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As a consequence of their close association with basalts in the field, it is inferred that both trachytes and basalts were derived from a common magmatic source: in fact, trachytes are regarded by many as being the ultimate products of differentiation of a world-enveloping basaltic magma.

(b) *Feldspathoidal Trachytes and related Rocks; Phonolites, Tinguaites and Leucitophyres*

Two groups may be distinguished according to the dominance of nepheline on the one hand, and leucite on the other. We term the former phonolites, and the latter, leucitophyres.

Phonolites are equivalent to nepheline-syenites among the coarse-grained rocks, and like the latter are typically felsic rocks in which feldspars, together with feldspathoids, are dominant over the coloured silicates. The distinctive feature is the occurrence of an alkali-feldspar, most typically sanidine, with a *sodic* feldspathoid, usually nepheline, and a subordinate amount of amphibole or pyroxene of NaFe-rich type, such as aegirine, riebeckite or the much rarer cossyrite.

In hand-specimens phonolites are very compact, grey-green rocks which are supposed to give a sonorous ring when struck by a hammer—hence their name, which is a classical rendering of an old term, “clinkstone,” used by A. G. Werner with the same significance. Texturally most phonolites are porphyritic, the phenocrysts including both nepheline and sanidine. A very well-known example which is found in most teaching collections is figured below (Fig. 99).

The quantity of nepheline in phonolites is very variable, and it is clearly necessary to distinguish between those in which the mineral has the status of an essential component, and those in which it is an accessory only. The latter are the connecting links between true phonolites and saturated trachytes. In the sense that they consist of the minerals appropriate to trachyte, with nepheline in addition, the logical name for them is nepheline-trachyte; but as Johanssen and others unfortunately have already used this term in another sense, the alternative of **phonolitic trachyte** may be used. An example familiar to British petrologists forms the Traprain Law Laccolith in Haddington, Scotland. The greater part of this rock consists of fluxionally arranged laths of alkali-feldspar, and the nepheline, which amounts to only 4 per cent, is elusive in thin sections; but it is supported, as it were, by 20 per cent of analcite and a few grains of sodalite.¹ Strictly the Traprain Law rock is a sodalite-bearing phonolitic analcite-trachyte.

Even closer to trachyte is a similar rock which forms a plug,

¹ Macgregor, A. G., *Geol. Mag.*, 59 (1922), p. 514.

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known as the Bass Rock, in the Firth of Forth, Scotland. Soda-orthoclase makes up 85 per cent of the rock, nepheline, sodalite and analcite all occur, but only in very small amounts, about 1 per cent, while the coloured silicates, aegirine-augite and iron-rich olivine, total some 8 per cent.

The phonolite which forms the Wolf Rock off the Cornish coast is much more typical than these Scottish rocks. In addition to

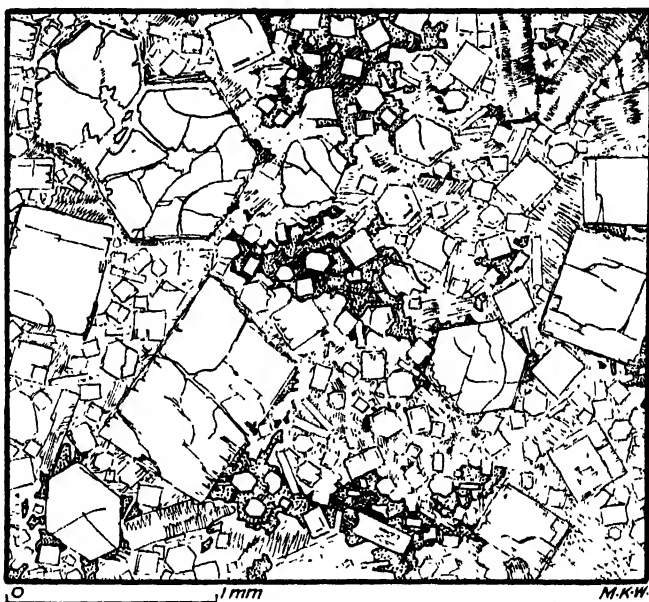


FIG. 99

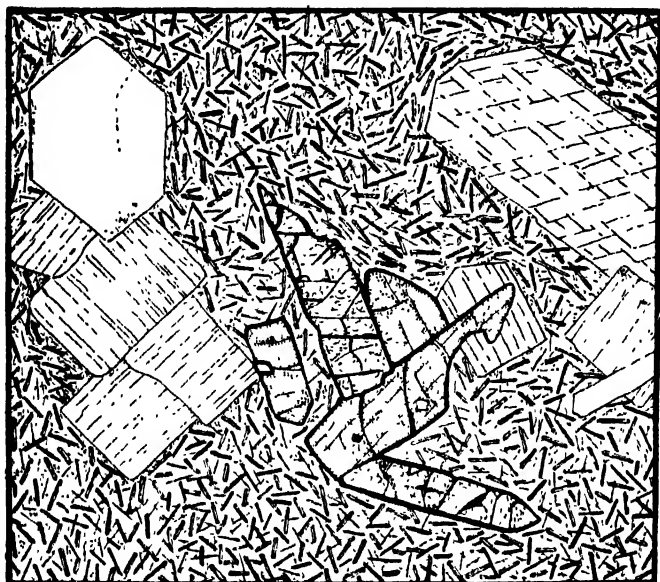
Phonolite, Brux, Bohemia. Both sanidine and nepheline occur in crystals of two generations. The former shows alteration parallel to the basal cleavage. The latter forms almost square or hexagonal sections, outlined in some cases by dark poikilitic riebeckite aggregates.

nepheline it contains much greyish nosean, and the term **nosean-phonolite** is applicable. Similarly, **sodalite-phonolite** contains all the minerals of typical phonolite, but in addition sufficient sodalite to be regarded as an essential constituent. The volcanic tract in Turkana, Kenya,¹ has provided a good example—a dense fine-grained green rock, which under the microscope is seen to be composed chiefly of laths of sanidine, grains of aegirine and small poikilitic patches of sodic amphibole, together with abundant though small (0.025 mm.) euhedral sodalites. The rock is, of course, the fine-grained equivalent of sodalite-syenite.

¹ Campbell Smith, W., *Quar. Journ. Geol. Soc.*, **94** (1938), p. 522.

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Leucite-phonolite should contain the minerals of true phonolite, with leucite in addition. The rocks of the Roman volcanic province provide examples. Leucite and aegirine-augite occur as phenocrysts, while the groundmass consists of soda-orthoclase, nepheline, aegirine-augite and sometimes sky-blue hauyne. With increasing



A K W

FIG. 100

Porphyritic aegirine-phonolite (Tingua type), Brazil. Phenocrysts of nepheline (with basal cleavage); sanidine, brilliant green aegirine, and sphene, set in a groundmass in which minute aegirine microlites are abundant.

leucite and decreasing nepheline, leucite-phonolite grades into leucitophyre, which ideally contains no nepheline.

Of rather special interest are the basic, phonolitic trachytes of Kenya type, named **Kenyte** by J. W. Gregory,¹ from the type-locality, Mount Kenya, Kenya Colony, but known to occur also at Mount Erebus, Antarctica.² The distinctive feature is the nature of the feldspar which occurs as large prisms, rhombic in cross-section, of anorthoclase. In shape, if not in composition, these appear to be identical with the feldspars of rhombic cross-section occurring in the

¹ Redescribed by W. Campbell Smith in *Bull. British Museum (Nat. Hist.)*, Vol. I, No. 1, p. 3.

² Prior, G. T., "National Antarctic Expedition, 1901-04," *Nat. Hist.*, i (1907), p. 101.

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Oslo larvikites and rhomb-porphyrries. In this and other respects, both chemical and mineralogical—the occurrence of small quantities of olivine and nepheline¹—kenyte closely resembles larvikite and laurdalite, and it is undoubtedly their fine-grained equivalent. Rhomb-porphyry is the medium- to fine-grained dyke-equivalent, and, as might be expected, occurs in association with the kenytes on Mount Kenya.

Falling in the phonolite group as defined above are certain dyke rocks commonly known as **tinguaite**s, so named by Rosenbusch (1887) after occurrences in the Tingua Mountains near Rio de Janeiro, Brazil, where they occur in dykes or as marginal facies of larger bodies of nepheline-syenite (Fig. 100). When the name was introduced it was considered necessary to use one name for a particular mineral assemblage occurring in the form of lava, and a different name for the *same* assemblage occurring in a dyke. As already argued, we consider this duplication of rock-names unwarranted. The principle is outmoded, though many petrologists still call intrusive phonolites, “tinguaite>s.”

As may be seen on comparing Fig. 100 with Fig. 99 phonolites of both Brux and Tingua types are distinctive rocks. They contain the same felsic minerals—sanidine and nepheline—but the dark minerals and consequently the textures are different. The riebeckite in the Brux phonolite forms the characteristic “mossy” micro-poikilitic patches, while the aegirine in the Tingua dyke-phonolite is microlitic.

Rocks described under the name **leucite-tinguaite** from the Tingua Mountains and Bearpaw Mountains fall into the group under consideration. One variety is an intrusive leucite-phonolite; others with more leucite may well be aegirine-leucitophyres. Their interest lies chiefly in the fact that they contain well-preserved pseudo-leucites which on analysis have been shown to consist of sanidine (66 per cent), nepheline 30 and acmite 3 per cent. The potassic feldspar and nepheline are, of course, the normal products of the breakdown of original Na-rich leucite under relatively slow cooling conditions, while the small amount of pyroxene doubtless represents mafic material adventitiously incorporated during the growth of the phenocrysts.²

Leucitophyres are the corresponding types characterized by the association of alkali-feldspar with leucite, together with mafic minerals in variety. Nepheline and nosean are common associates with the leucite, and as the feldspar is potassi-sodic and the mafic

¹ Campbell Smith, W., “Classification of some Rhyolites, Trachytes and Phonolites from part of Kenya Colony,” *Quar. Journ. Geol. Soc.*, **87** (1931), p. 242.

² Zeis, E. G., and Chayes, F., “Pseudoleucites in Tinguaita from Bearpaw Mountains,” *Journ. Petr.*, **1** (1960), pp. 86–98.

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minerals also contain some Na', the allocation of alkali atoms has been complicated indeed. The leucite usually forms large phenocrysts, distinctive both in hand-specimens and thin sections. The best-known rock of this type occurs as a dyke near Rieden in the Eifel. In the hand-specimen numbers of small phenocrysts of leucite

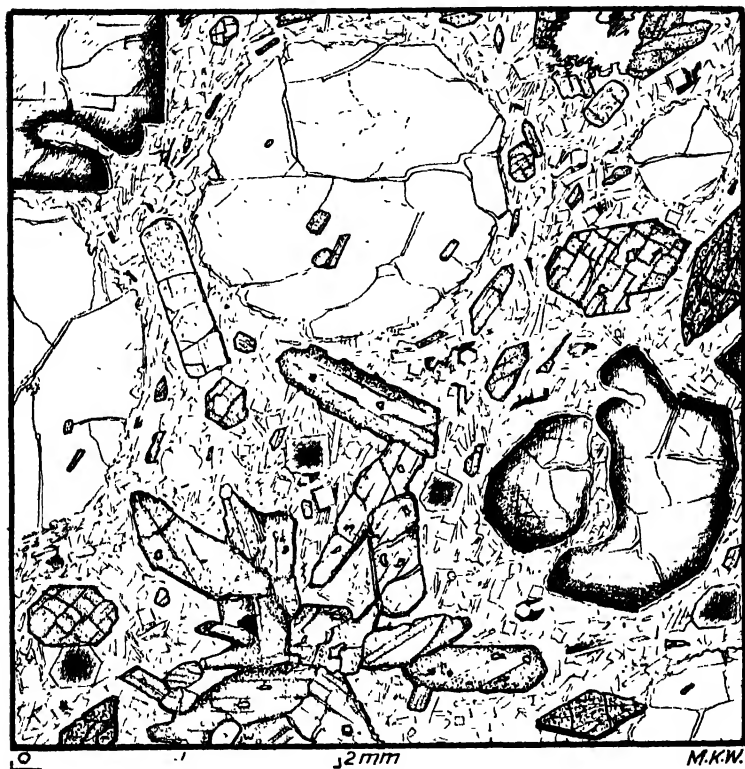


FIG. 101

Noscan-leucitophyre, Rieden, Eifel. Megaphenocrysts of leucite and nosean are about equally abundant, occurring with green-rimmed sodic pyroxene, set in a groundmass of sanidine and minute euhedral nepheline crystals. Apatite and sphene are important accessories.

and nosean are plentifully scattered in a greyish groundmass which also contains small black prisms of pyroxene. Under the microscope the phenocrysts of leucite and corroded noseans with heavy black borders are very striking (Fig. 101). Small nephelines of typical shape occur in the groundmass only, where they are associated with equally small laths of sanidine. Aegirine-augite is abundant, while sphene, apatite, magnetite, and less commonly melanite garnet are

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usual accessories. The proper designation of this well-known rock is nosean-nepheline-leucitophyre.

Leucitophyres occur together with other related types in the Leucite Hills, Wyoming. One variety consists of equal amounts of leucite, sanidine and coloured minerals. Among the latter, phlogopite, a rare mineral in normal igneous rocks, is dominant, forming small phenocrysts visible in the hand-specimen. It is accompanied by minute needles of diopside and a kataphorite-like amphibole. This is a type of **phlogopite-leucitophyre**, and for its silica percentage is particularly rich in potash.

Hauynophyres

Theoretically the fine-grained type, normally occurring as lava-flows and characterized by the assemblage hauyne-sanidine-coloured minerals should be distinguished as hauynophyre. With the incoming of leucite, hauynophyre grades through leucite-hauynophyre into hauyne-leucitophyre; and with the incoming of nepheline in increasing amounts it grades into phonolite *via* the intermediate type, nepheline-hauynophyre.

Summarily the fine-grained equivalents of under-saturated syenites consists fundamentally of two mineral associations:

- (1) alkali-feldspar with nepheline (and mafics) = phonolite;
- (2) alkali-feldspar with leucite (and mafics) = leucitophyre.

With the addition of another feldspathoid in an accessory capacity we have:

- (3) leucite-phonolite—alkali-feldspar, nepheline, (leucite), mafics;
- (4) nepheline-leucitophyre—alkali-feldspar, leucite, (nepheline), plus mafics.

These rock-types in a sense are partially under-saturated trachytes; but under-saturation is carried very much farther, indeed to the limit in some feldspathoidal lavas which contain *no feldspar*. Their nomenclature presents a difficult problem; but for the moment they may be referred to as feldspar-free feldspathoidal lavas, which is rather clumsy but true. In spite of their trachytic affinities these lavas include some of the most basic rocks in the world. There are, however, other feldspathoidal lavas which contain plagioclase and whose affinities are with the basalts as described in a later chapter. It will be a great convenience to consider all these lavas together under the general heading Feldspathoidal Lavas.

Origin of syenites and trachytes

In considering the nature and modes of origin of syenites and trachytes it is necessary in the first place to establish the relationship

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between them. To simplify matters as far as practicable we will exclude, for the moment, the feldspathoidal rocks within this range of compositions: they present a formidable problem of their own which can only profitably be considered when all the under-saturated (feldspathoidal) lavas and intrusives have been reviewed—at this stage only a few have been introduced.

We are concerned, therefore, with orthosyenites and orthotrachytes. So far as mineral composition is concerned their general equivalence has been established by the foregoing petrographic account. Such differences as do exist are understandable as arising directly from the different environments in which consolidation took place. High temperature mineral phases in the trachytes, for example sanidine and anorthoclase, are represented by orthoclase, microcline, albite and various perthites and antiperthites in the orthosyenites. It may safely be claimed that a trachyte can be found which will match very closely the composition of any given syenite. The trachyte is known to be a lava, *i.e.*, it is magmatic; the corresponding syenite cannot be proved to be comagmatic, though field relations sometimes strongly suggest that this is so. Given that trachytic magma is a reality, it follows that somewhere and sometimes trachytic magma must crystallize below the surface and syenite must result. Trachyte can be seen filling dyke-fissures which if traced downwards would surely link up with a syenitic rock body.

Demonstrably many syenites have not originated in this way: they seldom form bodies that are syenite all through—normally they are seen to form part, often only a small part, of much larger masses of granitic composition. The well-known Dresden syenite, for example, is marginal to the Meissen granite..

Again field relations coupled with laboratory studies have shown clearly that certain syenites were formed by *in situ* modification (desilication) of previously existing rocks. This process is illustrated by the outer zone of the Alnö Complex in Sweden, which is considered in detail in the discussion on the origin of the feldspathoidal rocks.

The converse process involving the acidification of more basic material has also occurred, demonstrably on a small scale, at Kiloran Bay, Colonsay, in the Hebrides. An ultramafic rock consisting largely of hornblende crystals contains numbers of quartzite xenoliths which are surrounded by thin rims of quartz-syenite. The latter has obviously resulted by interaction between the hornblendite and the siliceous inclusions. This may be a unique occurrence; but it does at least indicate a possibility.

There for the moment we leave the matter; but trachytes grade into nepheline-bearing phonolites and orthosyenites into nepheline-syenites and these are discussed later.

DIORITES, MICRODIORITES AND ANDESITES

THE predominance of alkali-feldspars and/or feldspathoids makes it relatively easy to define, classify and name syenitic rocks; when dealing with the plagioclase-dominant suite the matter is complicated by the fact that, by tradition, it is necessary to subdivide into two sub-groups: diorites and gabbros, together with their medium and fine-grained equivalents (microdiorites and andesites) on the one hand and microgabbros (dolerites) and basalts on the other. At different times and by different authorities the division has been made on the basis of the following criteria: (i) silica percentage; (ii) nature of the coloured minerals; (iii) kind of plagioclase present; and (iv) the percentage of coloured minerals—the colour index of the rock. Undoubtedly the easiest way of effecting the separation is to choose arbitrarily a convenient silica percentage, regardless of other considerations. This principle has already been discussed and found wanting; and for reasons already stated, we define rock-types in terms of their constituent mineral assemblages.

The position then is this: there are three *visible* criteria involved—(1) the kind of plagioclase, (2) the kind(s) of mafic constituents and (3) the amount of the latter, *i.e.*, the colour index. These features combine to give distinctiveness to the rocks concerned; but it is virtually impossible to reconcile the evidence afforded by all three variables. There will always be the odd specimen which conforms to the definition in two, but fails to do so in regard to the third (see below). The best, indeed the only chance of achieving unanimity in the solution of this problem—the most important of its kind, is to apply one criterion rigidly, but to allow reasonable elasticity in relation to the other two.

The precedent has been established by A. Johanssen, and adopted by many other American petrologists, of using the kinds of plagioclase as affording the most satisfactory means of distinction between the diorite-microdiorite-andesite suite on the one hand and the gabbro-microgabbro-basalt suite on the other. As the distinction is being made arbitrarily, it does not matter a great deal where the division is made: but in many diorites and andesites the plagioclase

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is within the oligoclase-andesine range, while typical gabbros and basalts contain labradorite or bytownite. Therefore the logical division to choose is the boundary between andesine and labradorite, *i.e.*, An₅₀.

Apart from the kinds of mafic constituents present in these rocks, the amounts are important factors in naming and classification. The difficulty of using colour index for differentiating between the two plagioclase-dominant suites is illustrated by the following facts. Shand in his textbook advocated drawing the division at 30 per cent; but S. E. Ellis¹ after a careful statistical study of the problem concluded that this figure was too low and suggested raising it to 40. The position as we see it is this: most diorites have a colour index below 40, most gabbros above this figure; but surely a rock of dioritic aspect, containing the right kind of plagioclase and mafic minerals, should not be classified as gabbro merely because its colour index is, say, 43.

	An ₁₀	An ₃₀	An ₅₀
0	OLIGOCLASITE		ANDESINITE
10			
20	Oligoclase-		Andesine-
30	LEUCODIORITE		LEUCODIORITE
40			
50	OLIGOCLASE-DIORITE		ANDESINE-DIORITE
60			
70	Oligoclase-		Andesine-
80	MELADIORITE		MELADIORITE
90			
100	HORNBLENDITE		

Common hornblende is the characteristic mafic mineral in diorites, though biotite, augite and hypersthene are also encountered. Olivine is normally absent. The mafic constituents typical of gabbros are pyroxene and olivine. Let it be imagined for a moment that diorite is defined by the mineral association common hornblende, andesine feldspar. In a large collection of specimens the proportions of these two minerals would be found to vary within the widest possible limits as suggested in the accompanying table. The same applies to the mineral association pyroxene (augite), labradorite, which defines ordinary gabbro.

¹ *Min. Mag.*, 28 (1948), p. 447.

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Rocks consisting of these pairs of minerals, then, cover the whole range of colour index variation, but not all of them are acceptable as diorites and gabbros: those varieties of extremely melanocratic and leucocratic nature are excluded as shown in the table. They are sensibly monomineralic, consisting essentially of the minerals named, together with a very small amount, limited, it is suggested, to 10 per cent of other minerals. That leaves diorites in the widest sense as having a colour index between 10 and 90—that is very different from Shand's limit of 30, or Ellis's of 40; but it is the lesser of two evils—it does group together all those rocks in the coarse-grain category which consist of the same two minerals, in varying proportions. Within the diorites so defined it is advisable to subdivide into three categories as explained under "Principles of Classification," and as shown in the table on p. 247. The grain-size limits are the same as those stated above for the quartz-rich rocks. For general purposes the division between microdiorite and andesite is drawn at the limit of unaided vision; that between diorite and microdiorite is drawn at a limiting grain diameter of 1 mm.

DIORITES

Fundamentally, as stated above, diorites consist of plagioclase within the oligoclase-andesine range, associated with mafic minerals and accessories, of which common hornblende, biotite, augite and hypersthene may represent the former, and sphene, apatite and magnetite the latter. Normally these rocks do not contain quartz, but in the more acid varieties accessory quartz may occur, usually in association with biotite, in **quartz-mica-diorites** (= tonalite).¹ There is little difference mineralogically between the latter and granodiorite: the same minerals are present in both types—only the quantities are different. Granodiorite grades into quartz-mica-diorite and little harm is done if the one is identified as the other because the distinction is based upon an arbitrarily chosen limiting percentage of quartz. The American practice in regard to the use of the two terms granodiorite and tonalite (quartz-mica-diorite) is somewhat different from the British. In **tonalite** as used by Johanssen, for example, alkali-feldspar is excluded and no limit is laid down as to the amount of quartz which may be present. Granodiorite, by implication, should share the characters of granite and diorite in the sense that both alkali-feldspar and plagioclase should occur, as in the rocks that we term adamellites. It must be admitted that the American practice is the more logical: it ensures that *all* rocks consisting of quartz, accessory *and* essential, mafics and plagioclase

¹ Cf. Bailey, E. B., who in the "Geology of Glencoe and Ben Nevis," *Mem. Geol. Surv.* (1916), p. 159, restricts tonalite to quartz-diorites containing hornblende and biotite.

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within the dioritic range are covered by the one name, tonalite. Our practice is to use two names one of which (granodiorite) is somewhat of a misnomer.

In the original tonalite from Monte Tonale in the Tyrol biotite accompanies hornblende. In Britain tonalites have been described

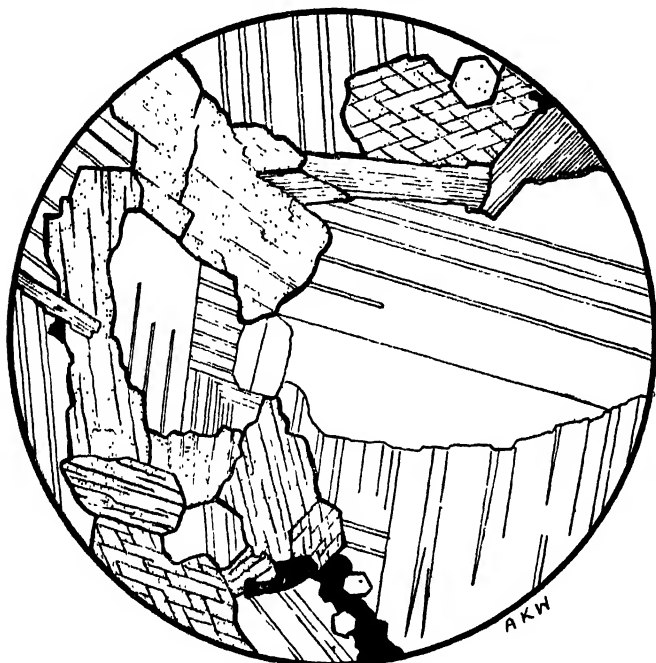


FIG. 102

Biotite-hornblende-diorite, Haute Savoie, France.

Plagioclase, common hornblende, biotite, magnetite and apatite shown. Drawn in plane polarized light, but the twinning of the plagioclase is indicated.

from the neighbourhood of Loch Awe, south-east of the Ben Cruachan granite.¹ The composition is as follows: plagioclase (An₃₀) 72, micropertite 11, quartz 7, dark minerals and accessories 11 per cent. Similar tonalites occur elsewhere in Scotland as integral parts of the Caledonian plutonic complexes, including those of Galloway.²

The diorite illustrated in Fig. 102 represents the ideal type: it is not over-saturated, biotite is subordinate to common green hornblende, the plagioclase, as is commonly the case, is zoned, with

¹ Nockolds, S. R., "The Contaminated Tonalites of Loch Awe," *Quar. Journ. Geol. Soc.*, **90** (1934), p. 302.

² Gardiner, C. I., and Reynolds, S. H., "The Loch Dee Complex," *Quar. Journ. Geol. Soc.*, **88** (1932), p. 1.

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the cores more calcic than the outer parts of the crystals. The normal accessory minerals liable to occur in diorites of all kinds include sphene, probably the most prominent, apatite and magnetite. In the more basic diorites the place of hornblende tends to be taken by either clinopyroxene which is colourless in thin section, or by orthopyroxene, usually identified as hypersthene. This mineral is the "trade-mark" of the Charnockite Series, and as noted above, charnockitic complexes include rocks in which the dominant feldspar is plagioclase. Quartz-poor or quartz-free **enderbites** (Tilley) are hypersthene-diorites in the wider sense, though their charnockitic affinities may be shown by their field relations.

As regards texture, there is little special to record: it is typically hypidiomorphic granular. Most teaching collections contain specimens of "**corsite**" or "**napoleonite**," the best known orbicular rock, which until recently was classified as orbicular diorite; but in view of its thoroughly basic character—it contains only 46 per cent of SiO_2 —and the calcic nature of the plagioclase, which is bytownite (An_{75-85}), it has been transferred to the gabbros. In respect of its coloured mineral (almost exclusively hornblende) and its low colour index (21) it is dioritic; but in respect of its feldspar and its chemical characters it is gabbroic. The orbs make up about half the rock and consist of alternate shells of bytownite and amphibole of a somewhat bladed habit. The matrix is normal.

Meladiorites differ from normal diorites only in their higher content of dark minerals. Examples have been described from the Glen Fyne–Garabal Hill Complex in South-West Scotland. They are much more extensively exposed on the coast of South-East Jersey¹ and northern Guernsey, Channel Islands. These rocks are certainly meladiorites in the general sense, but are distinctive among the latter by containing euhedral hornblende prisms, in the Channel Island examples, often with hollow cores, and fluxionally arranged. These **appinites** (E. B. Bailey, 1916) are often pegmatitic, of relatively coarse grain and occur either in "pockets" in normal diorite or in the upper flux-rich portions of differentiated dioritic sheets.²

In the field meladiorites grade into **hornblendites** by the elimination of plagioclase, and the (nearly) pure hornblende rock is the ultimate mafic differentiation product of the dioritic suite. The nature, origin and crystallization history of the ultramafites in general are considered in detail in connection with the gabbroic suite. They are

¹ Wells, A. K., and Bishop, A. Clive, "An appinitic facies associated with certain granites in Jersey, Channel Islands," *Quar. Journ. Geol. Soc.*, **111** (1955), p. 143.

² In Survey publications appinites are defined rather differently. Appinites are regarded as hornblendic rocks in which the feldspar may be either plagioclase or orthoclase. This cuts right across the classification adopted in this book: the *orthoclase-hornblende* combination is syenitic, and in our view rocks of this composition should not bear the same name as the dioritic assemblage, *plagioclase-hornblende*.

regarded as "accumulative" rocks, in the sense that they are believed to have been formed by the accumulation in one place of crystals that were precipitated elsewhere and sorted, so that ultimately a monomineralic mafic rock resulted. It is difficult to believe that this process can apply to monomineralic hornblendite, and one important occurrence suggest a different mode of origin. At Garabal Hill several different kinds of ultramafite occur, chief among them augite-peridotite (dominant olivine associated with augite) which grades into hornblende-pyroxenite consisting of augite, orthopyroxene and hornblende as the third, minor constituent. The pyroxenite in turn grades into a pure, brown hornblendite termed *davainite* by Wyllie and Scott in the original description of the complex. The hornblendite is of particularly coarse grain, the individual crystals being prisms up to 2 inches in length. There can be little doubt that this extraordinary rock represents a facies of the pyroxenite which has been thoroughly reconstituted so that the place of the two pyroxenes has been taken by a single amphibole. Presumably the process was metasomatic, though no agent more chemically active than water-vapour, to provide the necessary (OH), would be required. Finally, the ultramafites noted above form only part of the Garabal Hill Complex,¹ and in so far as energy was needed for the above conversion, it was doubtless provided by the later members of the complex.

Finally, certain very rare rocks may be admitted as having some dioritic affinities though they are far removed from the normal rocks of this series. Their interest is two-fold: they contain nepheline and the rare accessory, corundum; and they grade into marginal facies which are practically pure plagioclase rocks, the feldspar being within the normal dioritic range. The proportions of plagioclase, nepheline and corundum vary considerably in different types: in one, there is much more nepheline than oligoclase, while corundum is of accessory status only. In another, nepheline occurs in small amounts only, while andesine and corundum make up the rest of the rock.

Plumasite, described by Lawson (1901) from Plumas County, California, forms a wide dyke cutting peridotite, and is fundamentally an oligoclase-corundum rock. The corundums are light-bluish in colour, up to an inch in length, and are embedded in a matrix of white oligoclase. Very similar plumasites have been described from Natal² and the Transvaal, South Africa, where again they form dykes, and tend to be of pegmatitic facies—the

¹ Nockolds, S. R., "The Garabal Hill-Glen Fyne Igneous Complex," *Quar. Journ. Geol. Soc.*, **96** (1940), p. 451.

² du Toit, A. L., "Plumasite . . . from Natal," *Trans. Geol. Soc. S. Africa*, **21** (1918), p. 53.

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corundums may be several inches in length. The marginal facies of these African rocks may be almost pure oligoclase, while the corundum becomes an important constituent in the more central parts.

Only slightly different from plumasite is **dungannonite** (Adams and Barlow, 1900)¹ from Dungannon in the Haliburton-Bancroft area in Ontario, a locality well known for its nepheline-syenites. Dungannonite is, in effect, a plumasite with a rather more calcic plagioclase taking the place of oligoclase. Andesine makes up about three-quarters of the whole rock, and large corundums about half of the rest. White mica, often tending to sheath the corundums, biotite, and perhaps the most important accessory, nepheline, also occur in small quantities. Again, some phases of dungannonite are almost pure andesine-rocks (andesinites). These rocks are of special interest because in the first place they are *under-saturated* diorites; and secondly, except for corundum-syenites (described above), they are the only igneous rocks which contain corundum.

From the petrogenetic point of view these corundum-bearing rocks are of special interest, though the problem of their origin has not yet been solved. One factor involved is demonstrably desilication: in several instances the dyke-rocks carry quartz until they penetrate, or at least come into contact with peridotite (or serpentinite, which is the same thing chemically). At the point where quartz disappears, the wall-rock shows a gain in silica, and corundum appears in the dyke-rock. But the loss of silica by the dyke-rock is not the only factor involved as there is a considerable gain in alumina to be accounted for. Those who have made a special study of certain Russian corundum-bearing rocks believe that the alumina was introduced by volatiles.

Reference back to the table showing the mineralogical relationship between the various members of the dioritic series will remind the reader that at the one extreme of composition occur rocks consisting essentially of plagioclase only. In this sense these rocks may be grouped for the purposes of classification with anorthosites which are more basic. **Oligoclase** and **andesinite** were both mentioned in the paragraphs immediately above, and it will be realized that they must be regarded as unusual rocks, of very limited distribution, and probably formed by different processes from those involved in the genesis of labradorite and bytownite, the two normal varieties of anorthosite.

MICRODIORITES

The general name microdiorite has been chosen for rocks of dioritic composition, but of medium grain-size, that differ only in texture

¹ "Geology of the Haliburton-Bancroft area, Ontario," *Geol. Surv. Canada Mem.*, No. 6 (1910), p. 315.

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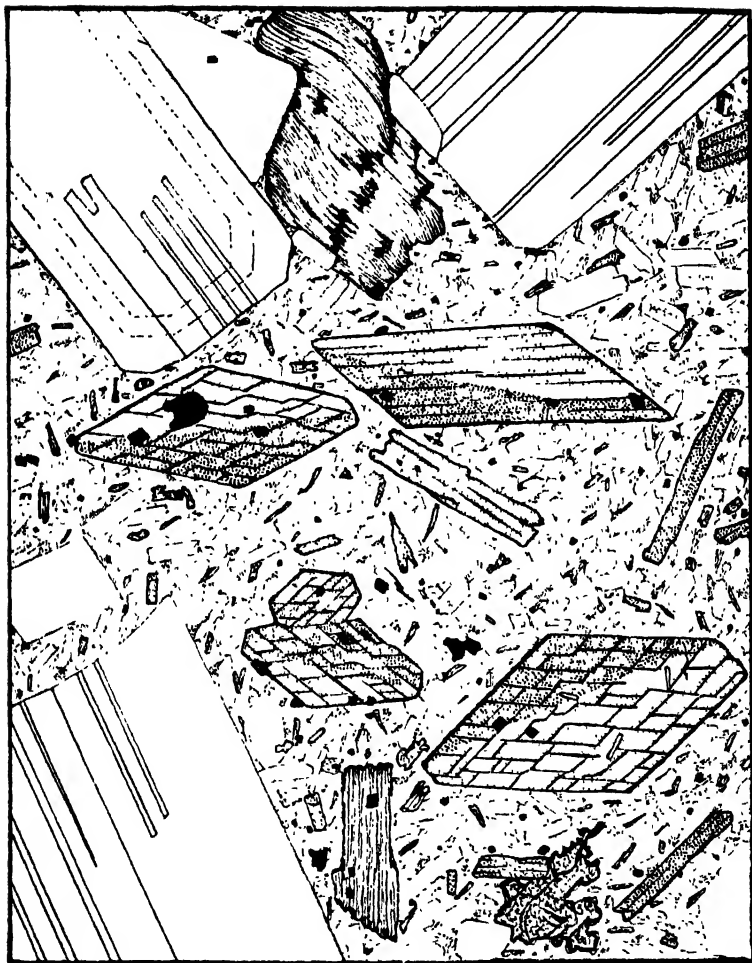


FIG. 103

Porphyritic microdiorite, shore section, Dalbeattie, Scotland.

Euhedral oligoclase-andesine, twinned and zoned, biotite distorted by pressure (top centre), common hornblende mostly twinned on (100), irregular sphene (bottom right centre) and granules of magnetite.

DIORITES, MICRODIORITES AND ANDESITES

from the diorites into which they grade. As the grain becomes finer, they pass into andesites. Though aphyric types are known, most microdiorites are porphyritic. To cover these textural variants two types are distinguished:

- (i) porphyritic microdiorites; and
- (ii) aphyric microdiorites.

The former are frequently named "diorite-porphyrries," but this name has been applied to porphyritic diorites, and is better not used. A synonym which is still widely used is "porphyrites"; but again there are many objections to its use, and in view of the several quite different usages, it is highly desirable to drop this term in favour of **porphyritic microdiorite**. In textural detail many of these latter rocks closely resemble andesites, and there is no doubt that many so-called "porphyrites" are as fine grained as andesites, and were named solely on account of their mode of occurrence as minor intrusions. Some of the "porphyrites" surrounding the Dalbeattie granite, for example, consist of a devitrified glassy matrix in which the abundant and relatively large phenocrysts are embedded. Although hypabyssal in mode of occurrence, such rocks are (intrusive) hornblende-biotite-andesites.

The nature of the dominant coloured mineral may be used for further subdivision, giving mica-microdiorite, hornblende-microdiorite, and less commonly augite- and hypersthene-microdiorites. Those specimens containing quartz (and usually biotite) are closely allied to tonalites and are to be distinguished as **microtonalites**. In such rocks quartz is restricted to the groundmass, where it may be intergrown with orthoclase as micropegmatite. Rocks of much the same general appearance but containing quartz phenocrysts in addition to those of other minerals prove on careful examination to be dacites, and in general this is a useful means of distinguishing between these two types. A rather special variety of over-saturated porphyritic microdiorite has been named **markfieldite**, from Markfield in the Charnwood Forest area in Leicestershire. A graphic intergrowth of quartz and alkali-feldspar forms the groundmass in which numerous phenocrysts (if they are so to be regarded) of plagioclase and hornblende are closely packed. As the feldspar is red stained, the general aspect of the rock is syenitic. An example of graphic microdiorite of a different type occurs at Penmaenmawr on the North Wales coast. In and around the Harlech Dome, also in North Wales, numerous minor intrusions, presumably of Ordovician age, occur. Many of these are of dioritic composition and some fall in the microdiorite category. They are noteworthy for the prominent phenocrysts of hornblende, plagioclase, and less commonly augite which they contain. Unfortunately they are not

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ANALYSES OF TONALITES, DIORITES AND MICRODIORITES

	I. Tonalite (Daly's average).	II. Quartz-diorite (Tonalite type), Carinthia.	III. Diorite (Daly's average).	IV. "Porphyrite," Glencoe.	V. Graphic microdiorite, Penmaenmawr type.	VI. Pyroxene- mica- diorite.
SiO ₂	62.35	63.09	59.67	65.30	58.45	54.07
Al ₂ O ₃	16.41	18.89	16.68	15.20	17.08	15.36
Fe ₂ O ₃	2.57	3.48	2.93	2.49	0.76	0.98
FeO	3.82	2.02	4.09	2.53	4.61	6.76
MgO	2.83	1.97	3.62	1.80	5.15	6.44
CaO	5.45	6.18	6.22	3.16	7.60	8.09
Na ₂ O	3.41	3.14	3.50	4.13	4.25	3.54
K ₂ O	2.13	1.30	2.13	3.37	1.02	1.96
H ₂ O	—	0.63	0.77	1.00	1.07	1.15
TiO ₂	0.67	—	0.39	0.83	—	1.30
Other constituents . .	0.36	—	—	0.23	—	0.38
	100.00	100.70	100.00	100.04	99.99	100.03

- I. Tonalite, Daly's average (1933).
 II. Quartz-diorite (Tonalite type), Wistra, Carinthia, Austria (W. Krezmar).
 III. Diorite, Daly's average (1933).
 IV. Porphyritic microdiorite, Glencoe, Scotland (Anal. E. G. Radley), *Ben Nevis Memoir* (1916), p. 183.
 V. Graphic microdiorite, Penmaenmawr, North Wales (J. A. Phillips).
 VI. Pyroxene-mica-diorite, Garabal Hill Complex, Scotland (Nockolds), *Q.J.G.S.*, 96 (1941), p. 451.

DIORITES, MICRODIORITES AND ANDESITES

particularly attractive subjects for petrographic study for they have experienced a mild regional metamorphism which has, in many instances, completely pseudomorphed the original minerals: hornblende is represented by aggregates of chlorite, epidote and calcite, while the plagioclase phenocrysts are now composed of paragonite, zoisite, etc.¹

Among the many "porphyrites" of the Scottish Survey officers, porphyritic and aphyric microdiorites occur in the dyke swarms related to the late-Caledonian and Old Red Sandstone volcanic centres in southern Scotland, for example in the Glencoe-Ben Nevis area, the Cheviot Hills, and associated with the granodiorite complexes of South-West Scotland (Galloway). They are in a much better state of preservation than the Ordovician Welsh rocks noted above, and include strongly porphyritic varieties similar to that illustrated in Fig. 102, containing phenocrysts of biotite, common hornblende and plagioclase (An₂₅₋₃₅). Many of these rocks are quartz-bearing and include microtonalite and the equivalents of the coarse-grained granodiorites, as is to be expected.²

ANDESITES

The fine-grained members of the Diorite Clan are collectively known as andesites. The essential distinction between andesites and microdiorites is the coarser grain of the latter, the division being drawn at the limit of unaided vision for the groundmass grains or microlites. Normally andesites thus defined occur as lava flows, but they may occur as minor intrusions, particularly dykes.

Andesites consist essentially of plagioclase within the oligoclase-andesine range. It is a refinement to distinguish between oligoclase-andesites and andesine-andesites: this can only be done on the basis of *normative* plagioclase on account of the impracticability of stating the bulk composition of the feldspar in any rocks containing both phenocrysts and groundmass microlites, which are inevitably different in composition. The matter is further complicated by zoning. Both oligoclase- and andesine-andesites may contain mafic minerals in variety: some are reputed to contain a little olivine, but augite, enstatite, hypersthene, hornblende and biotite are more typical. It is common practice to name specific andesites by attaching the name(s) of the dominant mafic mineral(s) before the general term "andesite," e.g., hornblende-andesite, hypersthene-andesite, etc.

¹ Wells, A. K., "Geology of Rhobell Fawr," *Quar. Journ. Geol. Soc.*, **81** (1925), p. 463; and summary account by the same writer in Matley, C. A., "The Harlech Dome . . .," *Quar. Journ. Geol. Soc.*, **102** (1946), p. 23.

² Phillips, W. J., "The minor intrusive suite associated with the Criffell-Dalbeattie Granodiorite Complex," *Proc. Geol. Assoc.*, **67** (1956), p. 103.

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Little is required by way of description of the minerals of which andesites are composed. The plagioclase phenocrysts are typically strongly zoned: probably the most striking examples of zoned feldspars occur in some of the Javanese and Japanese andesites. The amphibole in "hornblende"-andesites is usually the brown, basaltic variety (lamprobolite). Both it and the mica phenocrysts

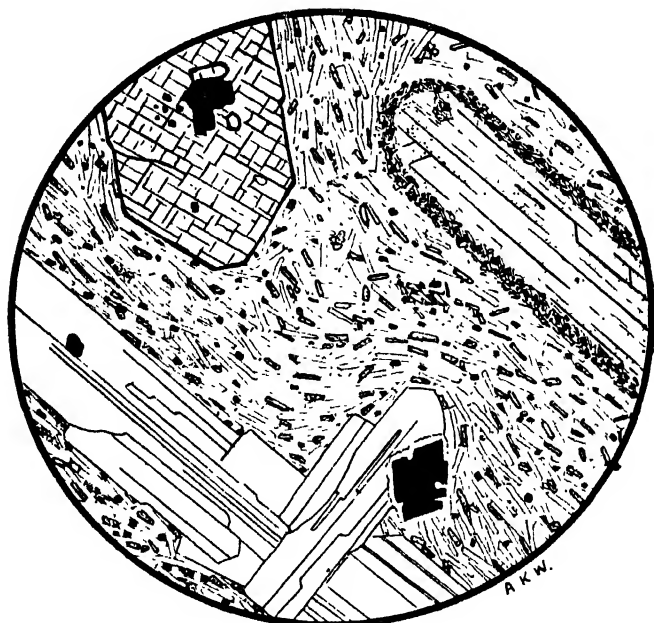


FIG. 104

Hornblende-augite-andesite, Col d'Entremont, Contal, Auvergne.

Complex twinned plagioclase, augite, and corroded hornblende phenocrysts are embedded in a groundmass of plagioclase microlites, together with prisms of colourless augite, and minute magnetite octahedra.

often show in the clearest manner that they were not in equilibrium with the magma in which they were carried to the surface. They may exhibit all degrees of magmatic corrosion. At an early stage this may amount to no more than a slight "peppering" with magnetite granules, but at a later stage of alteration the hornblende is progressively replaced by an aggregate consisting chiefly of granules of nearly colourless clinopyroxene and octahedra of magnetite (Fig. 104). In extreme cases the whole of the mica and hornblende may be so replaced, and only the shapes of the original phenocrysts

DIORITES, MICRODIORITES AND ANDESITES

survive. Finally, even this is lost, and an indefinite area rather richer in granular augite and magnetite than the general body of the rock is all that remains. In the same rock, phenocrysts of pale-coloured clinopyroxene may occur, and were apparently quite stable.

Magmatic corrosion and alteration of the phenocrysts in this way is probably due to reactions involving atmospheric oxygen

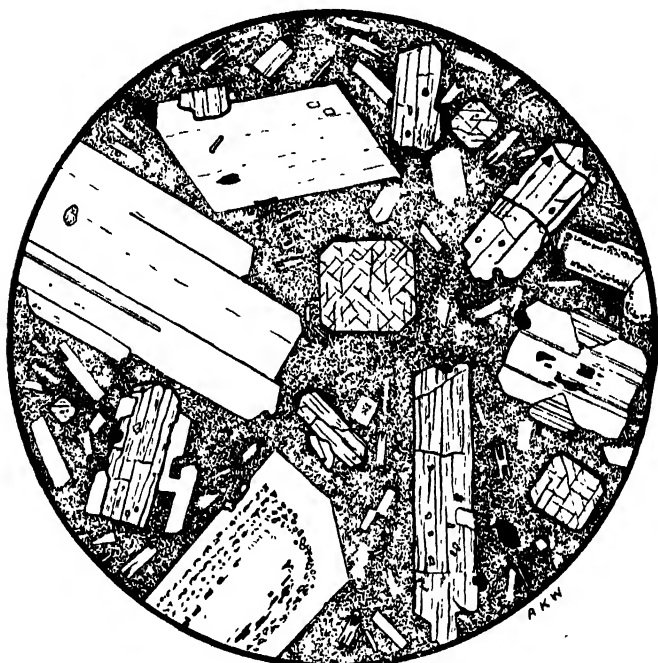


FIG. 105

Enstatite-andesite, Asama-yama, Japan.

Phenocrysts of euhedral zoned plagioclase and enstatite set in a subvitreous groundmass which contains a few skeletal feldspars.

which affect the lava when it reaches the surface, causing a temporary rise in temperature. Pyroxene replacing amphibole and biotite provides an interesting reversal of the normal sequence of discontinuous reactions of Bowen.

It is worthy of note that the clinopyroxene in typical andesites is a light greenish diopsidic variety, not the common brown augite so typical of basaltic rocks. The enstatite in andesites represents the olivine of more basic lavas, and is easily distinguished by its characteristic cross section—nearly square with the corners truncated (Fig. 105). Hypersthene is similar, but in thicker sections exhibits its

PETROLOGY OF THE IGNEOUS ROCKS

characteristic pleochroism. In the more ancient hypersthene-andesites the orthopyroxene is commonly replaced by the so-called bastite pseudomorphs.

The groundmass in which the phenocrysts are embedded may be wholly crystalline and composed largely of narrow microlites of plagioclase, associated, it may be, with granules of coloured minerals and not much iron ore (Fig. 104). If the microlites are so closely packed as to exclude glass altogether, the texture is said to be **pilotaxitic**; but in some instances wedges of glass lie between the microlites, giving the **hyalopilitic** texture. By increase in the proportion of glass to microlites such andesites grade into true andesitic glass. The latter is brown in thin section and may be perfectly transparent; but with increasing age it tends to lose its transparency through devitrification, just as with rhyolitic glass.

Andesites are somewhat prone to alteration through a variety of causes. Solfataric action in the neighbourhood of an active volcano, slight regional metamorphism or even atmospheric weathering tends to render the plagioclase turbid, to replace it by secondary albite charged with zoisite or epidote, while the coloured minerals are

ANALYSES OF DACITE AND ANDESITES

	I. Dacite, Lassen's Peak, California.	II. Hypersthene- andesite.	III Hornblende- andesite.	IV. Hornblende- andesite, Glencoe.
SiO ₂	68·72	59·48	61·12	61·49
Al ₂ O ₃	15·15	17·38	17·65	14·98
Fe ₂ O ₃	1·16	2·96	2·89	1·51
FeO	1·76	3·67	2·40	4·05
MgO	1·28	3·28	2·44	3·22
CaO	3·30	6·61	5·80	4·56
Na ₂ O	4·26	3·41	3·83	3·59
K ₂ O	2·78	1·64	1·72	2·80
H ₂ O	0·74	0·74	1·43	1·91
TiO ₂	0·31	0·48	0·42	0·96
Other constituents (chiefly CO ₂) . .	0·30	—	—	1·38
	99·76	100·00	100·00	100·45

- I. Dacite, East End of Chaos, Lassen's Peak, California (W. F. Hillebrand).
 II. Hypersthene-andesite, average of 20 analyses quoted by Daly (1933).
 III. Hornblende-andesite, average of 24 analyses quoted by Daly (1933).
 IV. Hornblende-andesite, Glencoe, Scotland (Anal. E. G. Radley), "Glencoe and Ben Nevis," *Mem. Geol. Surv.* (1916), p. 183.

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progressively replaced by chlorite, epidote, calcite, etc. Any vesicles in the rocks become filled with such minerals as calcite, chlorite, prehnite and other zeolites. It was to andesitic rocks in this altered condition that the name "porphyrite" was first applied. A little secondary quartz is liberated during such alteration: there is usually little doubt that it is secondary, on account of its mode of occurrence and its associates. In some andesites, however, a little primary quartz is to be expected on general grounds: the extrusive, fine-grained equivalent of tonalite is quartz-andesite. The quartz is restricted to the groundmass in such rocks; but as the amount of free silica increases they grade into dacites, which commonly contain first-generation quartz.

DISTRIBUTION AND ORIGIN OF ANDESITES AND DIORITES

Lack of uniformity in nomenclature and disagreement among petrologists as to the precise meaning of the terms "basalt" and "andesite" make it difficult to write convincingly about the distribution of andesites. The outstanding fact appears to be that, in respect of the role of andesites among the volcanic rocks, the continental regions of the Earth contrast strikingly with the Oceanic basins. Andesites which conform in *all* particulars with the definition given above, and which are accepted by all petrologists as being typical of their kind, are well represented in the volcanic regions of western North and South America—the Rockies, Sierra Nevada and the Andes—also in the island arcs including Java, Sumatra, Japan and the Aleutian Islands. In some of the volcanic tracts in these mountain belts andesites in variety, including augite-, hypersthene-, hornblende- and mica-bearing types, are overwhelmingly dominant; elsewhere they are closely associated with basalts, dacites, rhyolites and rhyodacites. They are typical "Intermediate" lavas with silica percentage varying from 59 for average hypersthene-andesite to 61 per cent in hornblende-andesite. Now andesites of these kinds are virtually unknown in the "oceanic" islands, where the dominant lavas are olivine-basalts.¹ Associated with the latter, however, are subordinate flows which have proved especially difficult to name and classify. They are as basic as many basalts (average silica percentage, 49); they contain some olivine, but the feldspar present is oligoclase, which is outside the range usually allotted to basalts. On this one criterion alone many petrologists, following the current American practice, classify these rocks as "oligoclase-andesites." Others, more impressed by their obvious basaltic affinities, refer to them as "oligoclase-basalts"; while they

¹ Macdonald, G. A., "Dissimilarity between Continental and Oceanic Rock Types," *Journ. Petr.*, 1 (1960), p. 172.

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have also been classified as andesitic basalts, basaltic andesites, trachybasalts and trachyandesites. In view of this confusion we are strongly of the opinion that they should be given a distinctive name that does not imply affinity with either basalts or andesites. The obvious choice among available names is "**mugearite**," largely on the grounds of priority. The type was defined unambiguously by A. Harker¹ in 1904; and there is no doubt that the rocks in question conform closely to his definition. We would stress the fact that they have no genetic affinity with the true ("continental") andesites, from which they differ in certain important chemical and mineralogical respects.

It has been suggested by Macdonald that the corresponding type containing andesine should also be given a distinctive name, and he has suggested "**hawaiite**" as being appropriate. We do not question its appropriateness; but whether it is really necessary is another matter. One name covers oligoclase- as well as andesine-bearing diorites and andesites, and with these as precedents, it may be suggested that "**mugearite**" might similarly cover oligoclase- and andesine-bearing varieties of the rocks in question. Mugearite, following Harker's original description, has been regarded as a fine-grained basic igneous rock, occurring as lava-flows and in minor intrusions (the original mugearites from Skye occur in composite sills). In mineral composition mugearites share the characters of basalts (in regard to mafic minerals) and andesites (in respect of the plagioclase present, which is within the dioritic (andesitic) range). It may be either oligoclase or andesine, and these mineral-names may be used as qualifiers, if considered necessary.

With mugearites and "**hawaiites**" excluded from the category of andesite, it is true to state that andesites are restricted to the continental regions: they do not occur in the oceanic basins. The boundary between the two is called the **andesite line**, which approximately coincides with the limits of the sialic crust—a fact that many regard as of considerable petrogenetic significance, otherwise it would have to be dismissed as a mere coincidence.

The origin of andesitic magma is one of the major current petrological problems. Again we are dealing with something which is incapable of direct proof—at best any conclusions reached must be speculative. Discussion has centred around two aspects of the problem: can andesitic magma be produced by normal differentiation processes from basaltic magma; and secondly, what is the significance of the limited distribution of true andesites to the circum-Pacific and other orogenic belts? As regards the first question, some andesites are so close to normal basalts in composition that

¹ Harker, A., "Tertiary Igneous Rocks of Skye," *Mem. Geol. Surv.* (1904), p. 264.

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there can be no doubt that the former *can* be produced by differentiation from basaltic magma; but one hesitates to accept this as the solution of the problem in view of the immense quantities of andesitic magma involved in the western mountain chains of North and South America. In any event, it appears unlikely that andesitic magma could be derived solely by differentiation of a basaltic parent (for example by fractional crystallization) because such differentiation is known to follow a different course. This can be demonstrated

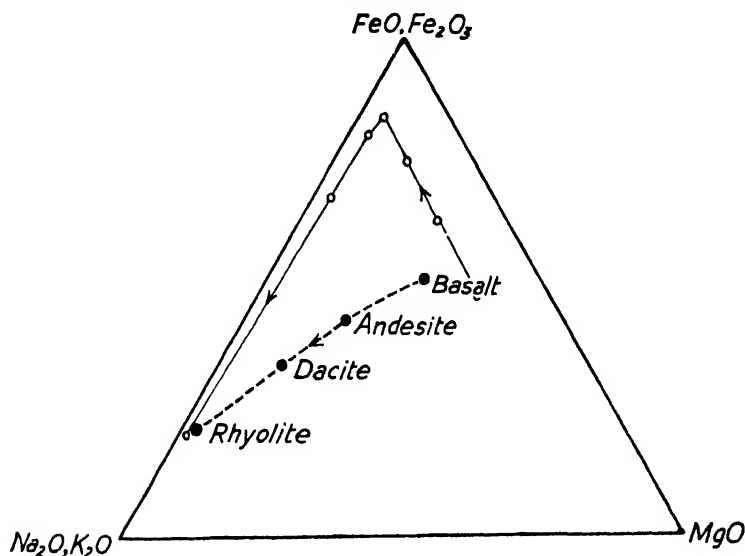


FIG. 106

Trend of differentiation due to fractional crystallization in the Skaergaard Complex (open circles) compared with the trend shown by the lavas belonging to the calc-alkali suite. Relevant oxide ratios for the latter are taken from average compositions quoted by R. A. Daly.

by plotting the composition of a suite of rocks known to have been produced by fractional crystallization differentiation (e.g. the successive layers of the Skaergaard Complex) on a triangular diagram together with common andesites as shown in the accompanying Fig. 106.

The restriction of true andesites to continental, orogenic regions strongly suggests that the evolution of andesitic magma is in some way related to the presence of "sial" in the crust. Wholesale acidification by reaction between basaltic magma and sialic rocks could provide andesitic magma in the required quantities, and in the present state of knowledge must be regarded as the most likely solution to the problem. The recipe for andesitic magma is probably:

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sial plus basaltic magma equals andesitic magma. In certain Tertiary igneous centres although true andesites are unrepresented, certain hybrids of approximately the same range of composition occur, for example in Arran. In this case the converse process appears to have operated: rhyolitic (granitic) *magma* has reacted with basic igneous *rocks* of earlier formation to form a local hybrid magma from which hybrid andesite, often intrusive, and hybrid diorite crystallized.

The composition of the deeper levels of the crust is so much an unknown factor that it is within the bounds of possibility that andesitic magma may be generated by melting of pre-existing crustal rocks in a mountain-root zone without involving either basaltic or rhyolitic magma in any way.

Turning now to the more parochial aspects of the distribution of andesites, in Britain they are well represented among the volcanics of two ages, the Ordovician and the Devonian. Andesites occur in the Arenig Mountain area of North Wales, but are more widely distributed in the Welsh Borderlands, where they build the Breidden Hills and form the Stapeley Volcanic Series of the Shelve area in Shropshire. In the English Lake District they are vastly more important, as the Borrowdale Volcanic Series, perhaps 10,000 feet thick, is dominantly andesitic, though basalts and subordinate rhyolites accompany the andesites. This is, therefore, a typical "continental assemblage." The same is true of the volcanic rocks of Devonian age in Scotland which are interbedded with sediments of continental facies (the Old Red Sandstone) and build various hill ranges including the Pentland, Sidlaw and Ochil Hills, the mountains of Ben Nevis and Glencoe, and the Cheviot Hills in the Border country. Andesites of Tertiary age, and of rather special types occur among the rocks forming the igneous complex of Central Arran: they were referred to above in the discussion on the origin of andesites. Further details, with references, will be found in the final part of the book.

The facts summarized above refer to true ("continental") andesites: mugearites occur in different areas and are of different ages. They form occasional flows interbedded with typical olivine-basalts associated with trachytic flows and minor intrusions among the Carboniferous igneous rocks of the Midland Valley of Scotland. They occur also as intrusions (composite sills) of identically the same petrographic type in Skye, but are Tertiary in age.

Origin and distribution of diorites

Diorites among coarse-grained rocks occupy the same position in the scheme of classification as andesites in the fine-grain category. It is tacitly assumed, therefore, that the two types are comagmatic; and that diorites result from deep-seated, slow cooling of andesitic

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magma. Certain facts of distribution and observable field relationships suggest that this is not necessarily true in all cases. Ordinary diorites, represented by hornblende-diorite, are uncommon rocks, whereas the assumed parental magma, as represented by andesitic lavas, is quantitatively second only to basalts. Further, in many instances diorites appear to have been formed by processes other than direct crystallization of andesitic magma.

All through the mountain ranges of western North and South America, *i.e.* in the regions where the dominant lavas are andesites, the associated coarse-grained types of major intrusion status tend to be quartz-bearing and often quartz-rich types (granodiorites, tonalites and monzonites). For example, in an area of some 4,000 square miles of the great southern California batholith,¹ over 2,500 square miles of surface area are occupied by tonalites and over 1,100 by granodiorite, these two rock-types together occupying over 90 per cent of the area studied. These are quartz-rich members of an assemblage of undoubtedly dioritic, *i.e.* andesitic affinities. Andesitic magma would crystallize directly to form such rocks as, say, hornblende-diorites only if it were undifferentiated; but the complexes we have in mind are differentiated, often strongly so, into divergent complementary rock-types. Those commonly exposed at ground level are the more acid products of differentiation, but the more mafic products are not unrepresented and it may confidently be asserted that they would become of increasing importance downwards.

In matters concerned with the status of diorites the role of microdiorites, especially the varieties commonly termed porphyrites, is important. Well-developed dyke-suites, covering the whole range of composition of the associated plutonic complexes, provide the essential link between extrusive andesites and deep-seated diorites. Although the fact has not been established on a statistical basis, one gets the impression that the rocks making the dyke-phase are closer to andesite than to the strongly differentiated granodiorite assemblage, in the sense of being less siliceous. The reader is referred to a recent account of the dyke-phase associated with the Criffel-Dalbeattie Complex, cited above, for fuller details.

To illustrate some of the general points referred to above we may use the Channel Islands, especially Jersey and Guernsey, where diorites are probably better developed than elsewhere in the British Isles. Jersey is best known as providing fine exposures of a wide variety of rhyolitic rock-types; but lying below these is a considerable thickness of more basic rocks, incompletely described as yet, known collectively as the "Jersey Andesites." Some of these contain

¹ Larsen, E. S., "Batholithic . . . Rocks of Corona . . . California," *Mem. Geol. Soc. Amer.*, 29 (1948).

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abundant platy plagioclase phenocrysts showing a well-developed fluxion structure, like that displayed by a well-known Scottish rock—the Carnethy Porphyry—from the Pentland Hills. Granites have broken into the base of this pile of andesites and have incorporated andesitic material forming various dioritic and monzonitic hybrids, many of them quartz-bearing. The granites contain xenoliths of all sizes. Some of them, embedded in alkali-granite, are rimmed with hybrid “diorite”; but internally the original structure has survived general reconstitution, and it is possible to recognize the fluxional, porphyritic andesite. Further, in this same complex certain diorites have demonstrably been formed by metasomatic acidification of gabbroic rocks: some of the hornblendes contain relics of the original pyroxene, while traces of the original ophitic texture have survived the metasomatism. Finally, in South-East Jersey, where dioritic rocks are extensively exposed along the coast, dioritic members of the complex appear to have been formed by direct crystallization from a magma of their own composition. They are frequently layered and often pegmatitic, containing hornblende prisms up to several inches in length.

Summarily, it appears that differentiation under deep-seated conditions causes andesitic magma to be represented by tonalitic and granodioritic rock-types rather than by ordinary diorites. Many of the latter have been produced locally by *in situ* recrystallization of material of appropriate (andesitic) composition; in other cases metasomatism during a wave of granitization has formed rocks of dioritic composition from pre-existing basic, gabbroic, rocks; but some, in the absence of evidence to the contrary, are believed to have originated by direct crystallization of a melt of their own composition.

This is a convenient place to consider some of the general aspects of hybridization and assimilation which are of such importance in the genesis of andesites and diorites.

Assimilation

Under this heading we shall consider some aspects of the problem of reciprocal reactions which may occur between magma on the one hand and intruded rocks on the other. The word “assimilation” has been used because it is easily understood and conveys a correct impression of what may happen in certain circumstances: rocks may become incorporated and completely digested or assimilated by magma so that no trace remains. However, assimilation is only one important end-product of the processes of magma/rock reactions to which the comprehensive term *syntexis* may be applied.

The first stage of the process of assimilation involves the fracturing of country-rock during a period of intrusion. Much of the shattering

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must result from the heating and differential expansion of the rocks in contact with the magma. Wall- or roof-rock which has been mechanically shattered and engulfed in the magma survives, for a time, in the form of angular blocks of all shapes and sizes, known as xenoliths. Initially the latter are clearly demarcated from the invading rock, and contrast with it in colour, texture and composition; but as a consequence of heating to magmatic or near-magmatic temperatures and soaking by magmatic fluids, they gradually merge their identity into that of the host rock. Two processes are involved in such assimilation. According to the composition of the xenolith, it may be incorporated by pure melting, or by reaction with the magma.

Bowen¹ has related the reactions between magma and xenolith to those which take place between a magma and its own normally precipitated crystal phases. Thus if a magma is temporarily in equilibrium with a plagioclase of composition, say, An_{50} , which is being precipitated, then the magma will *react* with an inclusion of a more calcic plagioclase (and therefore with a higher melting-point) until it conforms to the equilibrium conditions of the surrounding liquid. The latter contributes sodium and silicon, while the xenocrystic plagioclase yields up calcium and aluminium to the magma, until its composition is changed to that of the stable phase, *i.e.* An_{50} . On the other hand, any included plagioclase richer in sodium cannot exist at the magmatic temperature, which is above its own melting-point. Such a plagioclase must melt, and adds the whole of its substance to the magma. To melt a mineral, however, demands a relatively large amount of latent heat, which has to be provided by the magma. Since the latter probably has no surplus heat, unless it is being continually supplied by convectional transfer from depth, pure melting can only be achieved by a corresponding amount of crystallization—which releases latent heat.

What is true of plagioclase is equally true of all other groups of silicates belonging to continuously variable series, and probably true also of the discontinuous reaction series of Bowen (p. 168). Thus olivine, occurring in a xenolith immersed in magma precipitating biotite, would be progressively converted, in stages, from olivine to pyroxene, then to amphibole, and, if the reaction is not stopped by freezing of the magma, into biotite. Conversely, biotite immersed in olivine-precipitating magma and therefore at a temperature high above its own melting-point, would melt, and in so doing would cause the precipitation of an equivalent amount of olivine, necessary to provide the latent heat of fusion. In short, xenolithic minerals high in the reaction series tend to be "made

¹ Bowen, N. L., "The Behaviour of Inclusions in Igneous Rocks," *Journ. Geol.*, 30 (1922), p. 513.

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over" into those lower in the series, by reaction; while those low in the reaction series may be melted, on incorporation in magma. It follows from these general considerations that the individual component minerals in a xenolith of igneous rock will behave very differently, according to whether their melting-points are above or below the magmatic temperature at the time of incorporation in the magma.

All the more obvious effects of reaction between magma and country-rock are confined to the margins and roofs of intrusions. Xenoliths survive only where assimilation has been incomplete because of the exhaustion of the supply of heat energy of the magma. It is reasonable to infer that during the earlier history of an intrusion, and at greater depths, assimilation may not only have been complete, but it may have occurred very much more extensively than the evidence of a limited quantity of surviving xenoliths would suggest. Daly¹ described the process as abyssal stoping and assimilation, which he regarded as being of major importance, particularly as a means by which plutonic intrusions may make room for themselves as they invade higher levels of the crust.

Of course there are strict limitations as to the amount of rock that can be assimilated by any particular volume of magma. This depends primarily upon the heat available in the latter. It is generally accepted that magmas never have a surplus of heat over that needed to maintain them in a molten condition: in other words, they are never super-heated. Any work done by the magma in heating up and reacting with wall-rocks must therefore be at the expense of crystallization of some part of the magma. In some instances the heat available from this source must be very limited, and it is probably unwise to make too sweeping claims for syntexis, particularly in the highest levels of the crust. Here it would be necessary for magmas to use up much of their heat supply simply in raising the temperature of the relatively cold rocks. At greater depths any rocks involved in syntexis would be hotter to start with as a result of the geothermal gradient. Syntexis in depth may also be favoured by the kind of crustal compression and down-buckling which is believed to occur during orogeny. These are the conditions which are believed to have obtained during the formation of andesite-magma as discussed above.

It is always possible that similar end products may be produced by the mixing of *two magmas*, a process to which Harker gave the name of hybridization. As it is impossible to distinguish the effects of reactions between two liquids from those involving liquid and solid in most cases, the terms "hybrid" and "contaminated" are widely used as synonyms.

¹ Daly, R. A., *Igneous Rocks and the Depths of the Earth* (1933), especially Part II, Chapters xi and xii.

SYENODIORITES, MICROSYENODIORITES AND TRACHYANDESITES

ALL petrologists recognize the importance and significance of syenites and diorites, but there is not such unanimity concerning the need to distinguish the "middle series" which shares the characters of both. We believe that there is a place for these rocks of which the diagnostic characteristic is approximately equal status of alkali-feldspar (of syenites and trachytes) and oligoclase or andesine (typical of diorites and andesites).

The three rock-types named in the chapter heading are the coarse, medium and fine-grained members of the clan respectively.

SYENODIORITES (MONZONITES)

The type-rock was described (as monzonite) from the Monzoni Complex in the Tyrol. Admittedly monzonite makes up only a part of the whole complex, but this applies to many another valid type. The original monzonite is slightly over-saturated, with 2.5 per cent of free quartz, and therefore by analogy with syenites and diorites should be termed quartz-monzonite. Nearly two-thirds of the rock consists of feldspar, with andesine and orthoclase about equally balanced (32 and 30 per cent respectively). The coloured minerals include augite, biotite and hornblende in order of importance, the colour index amounting to 35.

The most significant fact in the mineral composition of monzonite is the coexistence of orthoclase and plagioclase, lying within the oligoclase-andesine range, as in diorites, in approximately equal quantities: neither must amount to less than one-third of the total feldspar. This is the crux of the definition, and is also the justification for using "syenodiorite" as synonymous with, and preferable to, monzonite. With increasing orthoclase syenodiorite grades into syenite, with increasing plagioclase into diorite. The amount of coloured silicates in monzonites so far described is relatively large compared with syenites, and doubtless melamonzonites will be discovered in due course, with colour index above 40; while feldspar-

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rich types, deficient in coloured minerals, would appropriately be termed leucomonzonites. No restriction need be laid down in regard to the kinds of coloured mineral present. They may be as in the type-rock, or any other combination appropriate to Intermediate rocks.

ANALYSES OF SELECTED TYPES OF MONZONITES

	I. Quartz- monzonite (Banat Type), Charnwood Forest.	II. Monzonite, Stratherrick	III. Monzonite, Monzoni.	IV. Monzonite, Glen Creran.	V. Monzonite, Brögger's Average.
SiO ₂ . .	64·30	59·34	54·20	53·22	54·90
Al ₂ O ₃ . .	17·89	18·91	15·73	17·20	17·44
Fe ₂ O ₃ . .	4·75	6·22	3·67	2·64	9·64
FeO . .	—	0·95	5·40	5·09	—
MgO . .	1·12	2·55	3·40	6·33	3·26
CaO . .	3·33	3·97	8·50	6·87	6·96
Na ₂ O . .	3·84	2·61	3·07	2·38	3·50
K ₂ O . .	3·37	2·91	4·42	4·12	3·51
H ₂ O . .	1·60	1·43	0·50	0·80	1·23
TiO ₂ . .	—	0·72	0·40	0·69	0·71
Other con- stituents	—	0·39	1·21	0·85	—
	100·20	100·00	100·50	100·19	—

- I. Quartz-monzonite, Banat Type, Croft Hill, Charnwood Forest, Leicestershire (E. E. Berry).
 II. Monzonite, Stratherrick, Glen Liath, Inverness-shire (W. Mackie), *Trans. Geol. Soc. Edin.*, viii (1905), p. 54.
 III. Monzonite, Monzoni, Tyrol (M. Schmelck).
 IV. Monzonite, Glen Creran, Scotland (Anal. E. G. Radley), "Ben Nevis and Glen-coc," *Memoir* (1916), p. 183.
 V. Brögger's average of fourteen typical monzonites. *Die Triadische Eruptionsfolge bei Predazzo*, p. 63.

In thin section there is usually no difficulty in recognizing the diagnostic feature of the monzonite type: the plagioclase tends to form comparatively smaller crystals of rather better shape than the orthoclase, which is poikilitic towards the plagioclase—a single plate enclosing, it may be, a large number of disorientated plagioclase laths as shown in Fig. 107. The pyroxene, when present, is an almost colourless diopsidic clinopyroxene, which in many cases is intimately associated with common green hornblende. The latter is irregularly moulded upon the pyroxene core as a rule, but the two may occur independently. Amphibole of a different kind—a somewhat fibrous "uralite"—tends to replace the pyroxene in altered monzonites.

SYENODIORITES AND THEIR EQUIVALENTS

Olivine-monzonites occur in the Oslo plutonic complex. Oligoclase and alkali-feldspar occur in approximately equal amounts. Although olivine is the most abundant coloured silicate in these rocks it is accompanied by clinopyroxene and biotite: the colour index is 35. The association of olivine with these kinds of feldspar is unusual, and olivine-monzonite of this type does not appear to

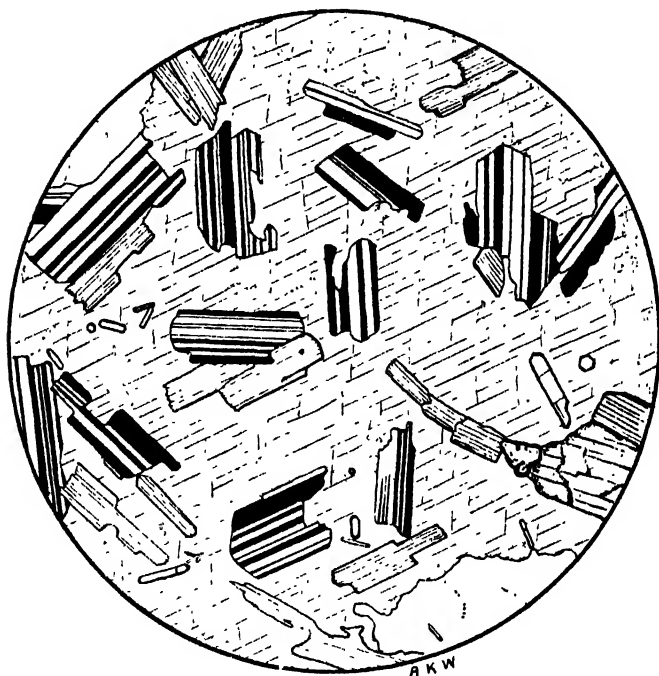


FIG. 107

Quartz-monzonite ("Banatite") above Pechadoire, Auvergne.

Subhedral plagioclase (twinning formalized), diopsidic augite (right), biotite (close ruled), apatite and sphene, all poikilitically enclosed in orthoclase making irregular contact with quartz at edge of section.

have been recorded elsewhere. Reference back to the description of laurvigite (larvikite) will show that although the list of minerals is the same in both rock-types, the proportions and the relationships between the feldspars are different.

Existing records suggest that monzonites are not common rocks; but this is partly a misconception, due to lack of uniformity in nomenclature: the rocks are there, but they have not been recorded as monzonites. As knowledge of the composition of the great "granitic" batholiths grows it becomes increasingly evident that

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they are only granitic in the widest sense. Quartz-bearing syenodiorites grading into granodiorites (American usage of the term = adamellites in the English sense) are quantitatively among the most important members in these great complexes. This is notably true of the Sierra Nevada batholith which is largely composed of rocks of these two types.

In Britain monzonites occur among the satellitic intrusions associated with the Devonian (O.R.S.) and Caledonian complexes in Scotland. They are represented also in the Tertiary Ring Complexes in Ardnamurchan, and although the dominant rock-types are basic—so-called eucrites and gabbros—the most central and latest intrusions consist of quartz-monzonites and tonalite.

As regards origin the view is widely held that syenodiorites are of hybrid origin in the sense discussed above under the origin of diorites.

A number of selected analyses are included in the appended table.

MICROSYENODIORITES (MICROMONZONITES)

For the sake of completeness we include, as the members of the medium grain-size group equivalent to the monzonites, rocks which, by analogy with the other clans, should be called micromonzonites. Under the name monzonite-porphyry (porphyritic micromonzonites) certain hypabyssal intrusives have been described from the well-known locality, the Henry Mountains in Utah, where they are associated with, and grade into, more abundant porphyritic microdiorites.¹

As noted above (p. 252), judged strictly on the basis of their feldspar-content, some Norwegian larvikites are monzonitic, while some rhomb-porphyrries—those of coarser grain—are technically porphyritic micromonzonites. The finer grained rhomb-porphyrries are trachyandesites.

TRACHYANDESITES

The fine-grained equivalents of the monzonites are the trachyandesites (Michel-Lévy, 1894), which bear to trachyte and andesite exactly the same relationship as syenodiorite does to syenite and diorite.

The meaning of the word trachyandesite is surely obvious: it applies to a rock possessing some of the characters of trachyte and some of those of andesite. A comparable group of more basic lavas

¹ Engel, C. G., "Igneous rocks . . . of the Henry Mountains, Utah," *Bull. Geol. Soc. Amer.*, 70 (1959), pp. 951-80.

SYENODIORITES AND THEIR EQUIVALENTS

is termed trachybasalt, and again the meaning is self-evident. Nevertheless the reader will encounter difficulty in connection with these rocks as a direct consequence of the different meanings of the words "andesite" and "basalt" as already explained.

In both groups of rocks alkali-feldspar, characteristic of trachytes, occurs (hence the "*trachy*-" part of the names). Both contain calc-alkali feldspar—plagioclase—in addition, and of the kinds appropriate to andesite and basalt respectively. Trachyandesites are characterized therefore by the association of alkali-feldspars and plagioclase (within the oligoclase-andesine range). Trachybasalts contain alkali-feldspar associated with more calcic plagioclase within the usual range labradorite-bytownite. Anorthite is not excluded, but is rare.

In typical trachyandesites the alkali-feldspar is usually sanidine: it may be rimming plagioclase phenocrysts but normally occurs as abundant, fluxionally arranged microlites in the groundmass. Sometimes the roles are reversed: the phenocrysts are sanidine, while the groundmass microlites are plagioclase.

Several named varieties of trachyandesites have been described, but many fall outside the group as defined in these pages. *Latite* (Ransome, 1898) was named after the Italian province, Latia, and has been adopted by Johanssen as the general name for the family; though surely "trachyandesite" is more informative than "latite."

Trachyandesites do not appear to be common rocks; but this may be partly due to the fact that not all petrologists recognize the monzonite (syenodiorite)—micromonzonite—trachyandesite suite, and therefore almost certainly trachyandesites have been recorded either as trachytes or as andesites. They do occur, however, in the Euganean Hills in Italy, among the lavas of the Siebengebirge and possibly among the Carboniferous volcanic rocks in southern Scotland, for example in Stirlingshire.

Trachyandesites are mineralogically equivalent to syenodiorites and the latter, as we have seen, are important in the so-called calc-alkali (continental) suite in orogenic zones such as the ranges of the Rocky Mountains, the Sierra Nevada and the Andes where andesites are the dominant types of lavas. Trachyandesites, too, occur in the same regions in association with andesites.

THE "BASIC" IGNEOUS ROCKS: GABBROS, DOLERITES AND BASALTS

Classification and nomenclature

THE one kind of magma which all petrologists are prepared to accept has the composition of olivine-basalt. This is potentially capable of crystallizing to a coarse-grained aggregate of plagioclase, pyroxene and olivine in a suitable environment. Natural rocks occurring as integral parts of the major layered basic intrusives contain these minerals in all possible proportions: individual rocks may consist of any one, two or three component minerals together, but the central types are gabbros and norites:

Gabbro consists fundamentally of plagioclase more calcic than An_{50} associated with clinopyroxene;

Norite consists essentially of plagioclase of the same range of composition associated with dominant orthopyroxene.

There is a complete gradation from gabbro devoid of hypersthene to norite devoid of augite. The one-pyroxene rocks we refer to as **orthogabbro** and **orthonorite** respectively. Between these extremes are the two-pyroxene members of the series including those which are fundamentally gabbros (though containing some hypersthene), and others, which are fundamentally norites, though containing some augite as shown in the appended table.

In addition to this line of variation, the rocks under discussion show a complete gradation between extremely leucocratic types on the one hand and extremely melanocratic varieties on the other. In the former, mafic minerals are virtually unrepresented, while in the latter, felsic minerals are practically absent. Thinking in terms of essential constituents, therefore, these extreme types are *mono-mineralic*, consisting in the first case of nothing but plagioclase, and in the second of pyroxene exclusively. They are, therefore, not covered by the definitions of "gabbro" and "norite" stated above, and special names must be used for them, as shown in the table. The intervening types, which are far more common, are adequately covered by the terms "gabbro" and "norite" coupled with suitable

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qualifiers. A rational scheme of nomenclature, based on the proportions of the two pyroxenes, and on the ratio of felsic to mafic components, is shown in the table which illustrates the relationships between the types mentioned above.

<i>Clinopyroxene dominant</i>	<i>Orthopyroxene dominant</i>
ANORTHOSITE	
LEUCOGABBRO	LEUCONORITE
GABBRO Hypersthene-gabbro	Augite-norite NORITE
MELAGABBRO	MELANORITE
CLINOPYROXENITE, e.g. diallagite	ORTHOPYROXENITE, e.g. bronzitite

Any single large basic complex, for example the Bushveld or Sudbury lopolith, might well provide a suite of specimens to which individually most of the names in the table might be applied: to illustrate this point in teaching, one of us uses a suite of specimens collected from the Bushveld Complex comprising anorthosite, leuconorite, norite, melanorite, pyroxenite, hypersthene-peridotite, dunite, chromitite and bronzite-chromitite, and titaniferous magnetite rock. All these are essential parts of this one rock-body, and were derived by differentiation from a common magmatic source; but an absurdly cacophonous term would result from hyphenating these nine rock-names in attempting to derive *one* name which would at least indicate the essential nature of the complex. The central type, representing the average most closely perhaps, is norite, and therefore we rather loosely use the term "Bushveld norite" for the whole rock-body.

It remains to note one other line of variation displayed by these rocks: as already noted, the plagioclase may be labradorite, bytownite or anorthite; and these mineral names may be added, as qualifiers, to the rock-type name when a high degree of precision is considered necessary. Thus distinction may be made between labradorite-gabbro, bytownite-gabbro and anorthite-gabbro. This applies also to noritic and troctolitic rocks of course. Labradorite is the normal gabbroic plagioclase and types containing anorthite are very rare.

Finally, it will be appreciated that all the rocks so far considered in this discussion are silica-saturated and may therefore carry accessory quartz. Others show various degrees of under-saturation

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by the presence of olivine. When the latter is subordinate in amount, the terms "**olivine-gabbro**" and "**olivine-norite**" should be used; but with increase in the proportion of olivine at the expense of pyroxene, a point is ultimately reached at which the latter disappears, and we are left with the two-mineral type, **troctolite**, consisting of olivine and plagioclase only.

Ideally (ortho) troctolite contains no pyroxene; but augite may progressively displace olivine to give continuous gradation from troctolite through varieties containing olivine *and* augite in varying proportions, ultimately into gabbro free from olivine. Similarly, orthopyroxene may displace olivine giving a gradation into olivine-free norite. Any one of these varieties may be represented by leucocratic, mesocratic or melanocratic types, according to the ratio of plagioclase to total mafic minerals.

GABBRO (Orthogabbro)		NORITE (Orthonorite)	
10	90	10	90
Olivine-gabbro		Olivine-norite	
30	70	30	70
Troctolitic gabbro		Troctolitic norite	
50	50	50	50
Gabbroic troctolite		Noritic troctolite	
70	30	70	30
Augite-troctolite		Bronzite-troctolite	
90	10	90	10
TROCTOLITE		TROCTOLITE	
100	0	100	0

The figures to the left and right refer to olivine, those in the middle column to pyroxene, indicating the percentages of the total mafics in both cases.

It will be noted that the naming of these rocks is consistent with the principles laid down elsewhere, and with the scheme used for the gabbro-norite series: thus, in the troctolite-gabbro range any rock containing *both* coloured minerals is named gabbro (with suitable modification) if augite is in excess, but troctolite if olivine is the more abundant. Similarly within the troctolite-norite range, so that, for example, a "troctolitic melanorite" is a coarse-grained aggregate of plagioclase, orthopyroxene and olivine, with the former in slight excess of the latter, while the total mafic silicates amount to 70 per cent or more.

Further, in most troctolitic intrusions, as in gabbroic ones, some form of internal differentiation, shown by layering, is prominent. There is a tendency for olivine to be concentrated towards the bottom of a layered unit, with plagioclase increasing towards the top. Therefore a gradation from monomineralic olivinite through

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melatroctolite, troctolite and leucotroctolite into anorthosite may be shown in extreme cases.

Finally, mention must be made of a basic plutonite termed **eucrite** in Survey publications. Properly this is the name of a type of meteorite and should not have been applied to a rock. The only rocks to which the name is applied occur among the Tertiary basic-ultrabasic complexes in Scotland. They are of mixed composition in the sense that they are neither clearly defined gabbros nor norites but combine the mineralogical features of both. In addition to olivine, two pyroxenes, hypersthene and augite, occur in approximately equal amounts, while the plagioclase is exceptionally calcic, being bytownite or anorthite.

Since the name was originally introduced (as *eukrite*) rocks of widely differing composition have been included under this name, and it is quite clear that the rocks in question could be covered by the terms "hypersthene-gabbro" or "augite-norite" in specific instances. We thoroughly agree with Le Bas's conclusion that this is one of the rock-names that serves no useful purpose and tends to confuse, rather than clarify the issue.¹

To indicate the interrelationships between the many rock-types resulting it would be necessary to use a tetrahedron, with plagioclase, olivine, augite and hypersthene at the four corners; but this cannot be adequately shown on a two-dimensional diagram.

In the detailed account which follows, the gabbros, norites and troctolites are considered together with their medium and fine-grained equivalents; while the products of extreme differentiation including the several monomineralic types, together with the phenomena of layering, are considered in a later chapter.

I. GABBROS, NORITES AND TROCTOLITES

After the above detailed discussion of the classification and nomenclature no difficulty should be experienced in identifying the ordinary gabbroic rocks. There are, however, certain points of detail which arise in the examination of thin sections to which we wish to direct attention.

Mineral composition of gabbros and norites

Plagioclase is normally somewhat in excess of mafic minerals in most gabbros and norites, amounting to about 60 per cent of the whole. Broad albite-twin lamellae are characteristic, combined in many cases with Carlsbad and/or Pericline twinning. The crystals tend to be of platy habit, flattened parallel to the side-pinacoid (010), and consequently in layered intrusions some degree of parallel orientation is commonly observed. The plagioclases are frequently

¹ Le Bas, M. J., "The term eucrite," *Geol. Mag.*, **96** (1959), pp. 497-502.

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schillerized. In some instances low-power magnification reveals the cause of the phenomenon: minute rods of iron-ore in parallel orientation. In hand-specimens these inclusions may render the crystals quite dark, and in some instances cause the familiar "play of colour." The rods often change their direction at twin planes, and may be orientated in more than one direction, causing a fine-grained graticule. When these inclusions are exceptionally small they may cause a general cloudiness which is irresolvable except under the highest magnifications. In certain circumstances—low-grade regional metamorphism or late-stage (deuteric) alteration—the plagioclase breaks down to an aggregate of Ca"-rich minerals including typically zoisite or epidote, and calcite, set in a matrix of secondary albite.

(Clinopyroxene is commonly augite) (the variety diallage); but in some cases is the lilac-tinted, slightly pleochroic titanaugite. This also may be schillerized in much the same fashion as the plagioclase; but a special point to look for is the presence of regularly orientated sheet- or film-like inclusions of orthopyroxene. (Similarly the orthopyroxene, usually hypersthene or bronzite,) may show inclusions of clinopyroxene, easily identified (when the lamellae are thick enough to produce an optical reaction) by the stronger birefringence of the augite as compared with the host mineral. These are features of exsolution as described previously under "Crystallization of Pyroxenes."

Under conditions involving incipient metamorphism pyroxene alters in one of two ways: it may change to amphibole or chlorite. The former type of change is often termed *uralitization*, though the amphibole commonly present is a light-coloured cummingtonite of fibrous habit, which develops first along the cleavages and round the periphery of the crystal grains.

(Olivine is a safe mineral to identify even though only a grain or two may occur in a complete thin section.¹ It displays no special features; but it may be perfectly fresh or replaced to any degree by serpentine, iddingsite or bowlingite. Although on account of the high temperature at which it crystallizes olivine is an early silicate to separate from a basaltic magma, euhedral crystals are seldom seen in these coarse-grained rocks on account of crystal-liquid reactions during slow cooling.

Accessories are few in number and generally limited to iron-ore, often identified (erroneously) as magnetite. In most instances it is a combination of ilmenomagnetite and ilmenite forming complex aggregates.

¹ These may easily be overlooked unless the observer takes the precaution of examining the *whole slide* with the naked eye or a low-power lens *before* placing it on the stage of the microscope.

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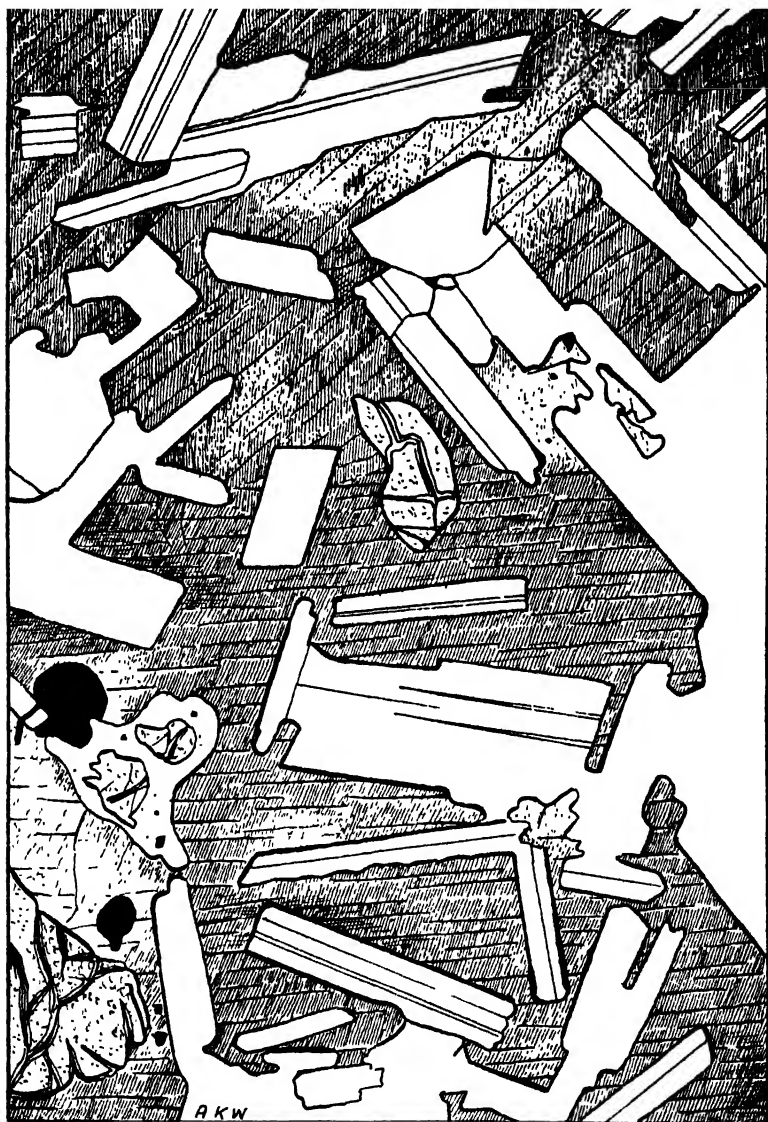


FIG. 108

Olivine-gabbro with ophitic texture, Kofayi, Nigeria.

Three small olivines partly altered to serpentine embedded in strongly schlierized clinopyroxene in ophitic relationship to labradorite. Iron-ore rare, in rounded patches and minute octahedra in the serpentine. (Size of section 6 mm. \times 9 mm.)

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Textural range. (Many gabbros and norites show the ordinary xenomorphic granular texture, typical of coarse-grained rocks in general; but some develop a very distinctive relationship between plagioclase and pyroxene, termed *ophitic texture*. When well developed, ophitic texture involves the enclosure of euhedral, disorientated plagioclase "laths" in extensive "plates" of augite (as shown in Fig. 108). This texture is by no means restricted to gabbros; if anything, it is more characteristic of dolerites and is therefore discussed more fully under that heading.)

(Orbicular norites and gabbros have been described from several localities) including Romsaas in South Norway, the Kenora District in Ontario, San Diego County, California¹ and Corsica. Under the name "corsite" (Zirkel) or "napoleonite" the rock from the last-named locality is probably the best known of all orbicular rocks, though it is often referred to as orbicular diorite, on the grounds that it consists of amphibole and plagioclase. Actually it is thoroughly basic,² with 46 per cent of silica; the plagioclase is bytownite (An_{75}), while the amphibole is, in part, paramorphic after pyroxene. Approximately three-quarters of the rock consists of bytownite and one-quarter of amphibole. The orbs average an inch in diameter, though they may reach three inches, and a central section shows a core of normal texture, surrounded by alternate shells of plagioclase and amphibole. The matrix is of normal texture.

(In olivine-gabbros the reaction relationship between the various coloured components is often particularly well displayed. Olivine is rimmed with pyroxene, amphibole envelops the latter, and in turn is surrounded by biotite (Fig. 109). The order of the successive rims is that of the minerals comprising Bowen's discontinuous reaction series. Reaction rims of a quite different nature provide one of the most fascinating textural characters of the igneous rocks. They are termed "coronas," and take the form of narrow mantles often of singularly uniform width, sometimes single, but in other cases double, round olivine crystals embedded in plagioclase. These textures therefore tend to be more uniformly developed in rocks of troctolitic type, and are more fully described under that heading.)

A characteristic feature of certain noritic rocks is the development of **symplectic intergrowths** along intercrystal boundaries. They consist of a myrmekite-like bulbous outgrowth of plagioclase, occasionally, though rarely, twinned in a normal manner, riddled with vermicular inclusions of orthopyroxene. In the example illustrated in Fig 127 the vermicules have grown away from

¹ Schaller, W. T., "Mineralogical Notes," Series 1, *Bull. U.S. Geol. Surv.*, No. 490 (1911), p. 58.

² Tröger, E., "Quantitative Daten einiger magmatische Gesteine," *Tsch. M.P.M.*, 46 (1934), p. 167.

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vein-like masses of titaniferous magnetite rimmed against labradorite with grains of fayalitic olivine associated at one point with poikilitic biotite. The olivine itself is locally intergrown with vermicular magnetite. There is no doubt that this phenomenon dates from a



FIG. 109

"Olivine-gabbro" (bojitic), Wolf Cave, Jersey.
Olivine, titanite, barkevikite and plagioclase.

Note the irregular fretted outlines of the barkevikite against the titanite; evidently the former is paramorphic after the latter. Actually in this field barkevikite is more abundant than pyroxene, and the rock is bojitic rather than gabbroic.

(Reproduced by courtesy of the Council of the Geologists' Association.)

late stage in the crystallization of the magma. Similar intergrowths of the same two minerals may be seen in the well-known European norite from Hitteroe, Norway (Fig. 110).

The foregoing description of mineralogical and textural features relates principally to gabbros and to a lesser extent to norites. The latter are generally of less importance than the former. Troctolites are sufficiently distinctive from the gabbroic rocks in several respects as to merit separate description.

Troctolite.—A typical specimen of troctolite, especially if somewhat weathered, is a striking-looking rock, the grey plagioclase-aggregate being studded with black, brown or reddish olivines or pseudomorphs after olivine. This accounts for the popular name,

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"troutstone," often applied to this rock. In many cases the olivines are seen to be insulated from the plagioclase by reaction rims of orthopyroxene, usually very thin, but in special types they may be thicker and double. If the olivine has been serpentinized, the expansion resulting from the change in composition causes intense



FIG. 110

"Norite," Hitteroe, Norway.

A large apatite is in contact with light green clinopyroxene associated with iron-ore. The pyroxene is fringed with a myrmekite-like intergrowth of the same pyroxene and plagioclase, lobed into schillerized labradorite. The latter contains regularly orientated short rods of iron-ore, and also colourless tubular inclusions whose orientation is controlled by the twinning. (Bottom, right.)

shattering of the surrounding plagioclase. The fractures radiate out from the olivine nuclei (Fig. 111).

The finest display of troctolites in Britain is found in the basic complex at Belhelvie in Aberdeenshire.¹ This is a layered concordant intrusion, partly ultrabasic, though mainly consisting of troctolite

¹ Stewart, F. H., "The Gabbroic Complex of Belhelvie, Aberdeenshire," *Quar. Journ. Geol. Soc.*, 102 (1947), p. 465.

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grading into norite, hypersthene-gabbro, and even, if one considers a small enough specimen, anorthosite and olivinite. In troctolites proper the proportion of plagioclase (An_{70}) to olivine (Fo_{76}) varies considerably. In an average specimen, the analysis of which is given under No. 11 on p. 317, the mineral composition is bytownite 70.5, olivine 28.4, and pyroxene 0.6 per cent.

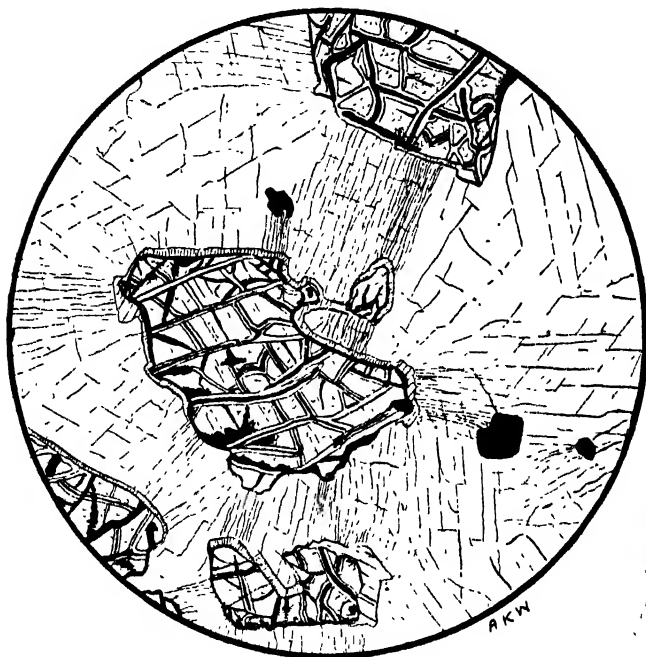


FIG. 111

Troctolite, Belhelvie, Aberdeen, Scotland. Olivines, veined by serpentine with separation of magnetite, in plagioclase traversed by systems of closely spaced fractures. Occasional primary magnetite crystals. A thin veneer of orthopyroxene surrounds the olivines which measure 2-3 mm.

Texturally the Belhelvie troctolites are interesting on several counts. The olivines are uniformly rimmed with reaction products, the inner zone consisting of orthopyroxene or amphibole in different cases—both colourless varieties—while the outer zone is a hornblende-spinel symplectite. Corona structure of this type is illustrated in Fig. 112, and is believed to be due to the action of liquid residua at high temperatures. Secondly, the degree of idiomorphism exhibited by the two chief minerals depends upon their relative proportions: with olivine in excess of a certain amount, this mineral separates before the plagioclase, and is therefore euhedral towards

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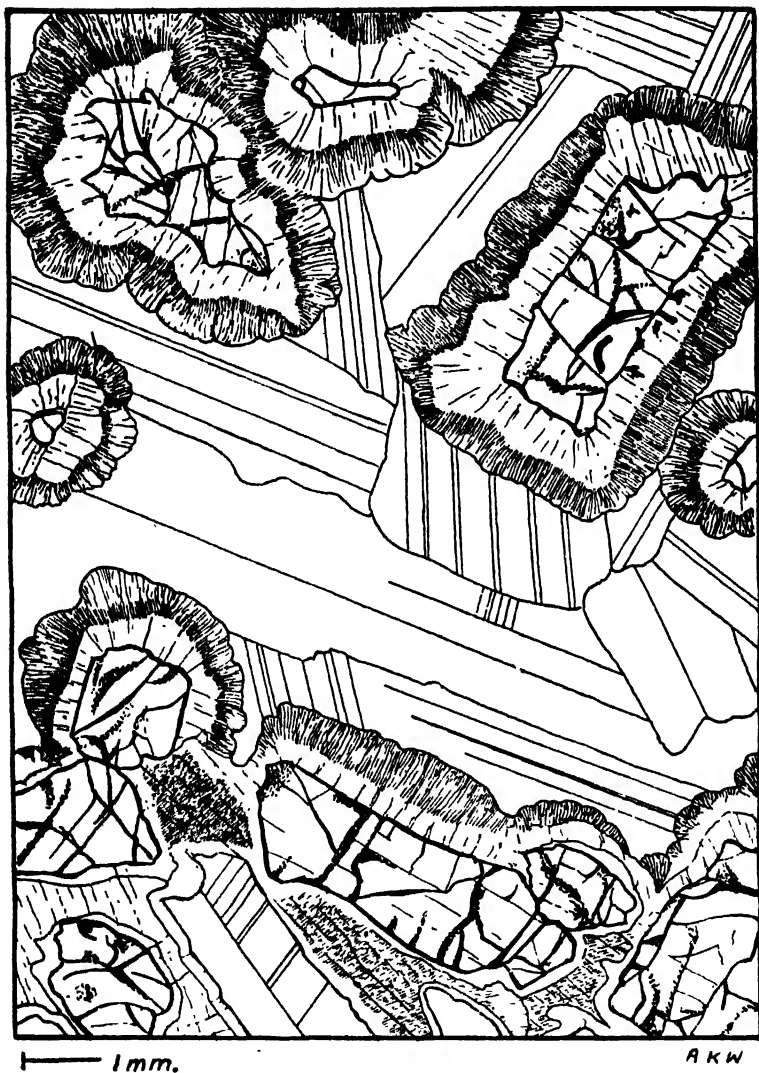


FIG. 112

Augite-cummingtonite-olivine-norite, Risør, Norway.

Corona structure is finely developed around cores of olivines. The coronas consist of orthopyroxene with a sharp boundary against an amphibole which in places is cummingtonite, but elsewhere the place of the latter is taken by a very light green symplectite, crowded with irregular vermicules. Cummingtonite occurs in association with bright green spinel around iron-ore, while the former occurs as a very thin but regular zone around schillerized clinopyroxene. (Lower part of the drawing.)

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ANALYSES OF GABBROS, NORITES, ETC.

	I. Allvalite, Rhum.	II Troctolite, Belhelvie	III. Marginal gabbro, Skaergaard.	IV Hypersthene- olivine-gabbro, Skaergaard.	V. Ferrohorthonolite- ferrogabbro, Skaergaard.	VI. Norite, Huntly.
SiO ₂	45.56	40.21	48.01	45.48	44.61	49.18
Al ₂ O ₃	21.17	24.22	19.11	16.41	11.70	16.00
Fe ₂ O ₃	1.10	0.54	1.20	2.09	2.05	0.02
FeO	5.59	4.48	8.44	9.29	22.68	8.22
MgO	11.48	9.30	7.72	11.65	1.71	9.47
CaO	11.42	11.41	10.33	10.46	8.71	12.54
Na ₂ O	1.99	1.92	2.34	2.06	2.95	2.04
K ₂ O	0.16	0.16	0.17	0.27	0.35	0.26
H ₂ O	1.28	6.86	0.60	1.03	0.42	0.48
TiO ₂	0.40	0.10	1.51	0.94	2.43	1.24
Other constituents	0.14	1.03	0.82	0.11	2.54	0.85
	100.29	100.23	100.26	99.79	100.15	100.30

- I. Allvalite, Allval. Rhum (Anal. M. Brown) *Phil. Trans. Roy. Soc.*, 240 (1956), p. 47.
- II. Troctolite, Belhelvie, Aberdeenshire (Anal. F. Stewart) *Quar. Journ. Geol. Soc.*, 102 (1947), p. 474.
- III. Olivine-gabbro, chilled marginal facies, Skaergaard (Anal. W. A. Deer), Skaergaard Mem., *op. cit.*, p. 140.
- IV. Olivine-hypersthene-gabbro, base of layered series, Skaergaard (Anal. W. A. Deer). *op. cit.*, p. 92.
- V. Ferrohorthonolite-ferrogabbro, Skaergaard (Anal. W. A. Deer), *op. cit.*, p. 106.
- VI. Norite, Huntly, Aberdeenshire (Anal. E. G. Radley), Huntly, *Mem. Geol. Surv.* (1923), p. 115.

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it. With plagioclase in excess, the olivine crystallizes late, and is interstitial towards the earlier, and therefore better formed, plagioclase. When present in eutectic proportions both crystallize together, neither has the advantage over the other, and neither therefore is euhedral. Finally, in places troctolites develop a pegmatitic facies, with crystals of olivine up to 2 cm. in diameter.

RELATIONSHIPS AND ORIGINS

In a sense it is premature to discuss these matters before the microgabbros and basalts have been considered; but there are certain aspects that may profitably be discussed at this stage. The general equivalence of gabbros, dolerites and basalts is a fact accepted by petrologists, though with some reservations concerning gabbros. The argument used in connection with the relationship between rhyolites and granites, trachytes and syenites, phonolites and nepheline-syenites, and andesites and diorites applies with even greater force to basalts and gabbros. Basaltic magma is the most widespread of all types, and basaltic dykes grade into dolerites—magmatically these two are one and the same thing: no petrogenetic significance attaches to the difference in grain-size. Basaltic and doleritic dyke-swarms constitute an impressive aspect of volcanism. If the individual dykes could be traced *downwards* they *must* emanate from a now congealed magma of the same composition but comprising equivalent rocks of grain-size appropriate to the depth at which they crystallized and the duration of their cooling history. This is a statement of fact; but whether it applies to all gabbros is another matter. There are good grounds for believing that syenites, nepheline-syenites and diorites have originated in more than one way, and this may be true of gabbros also. In particular we have already accepted the principle that in an appropriate environment a fine-grained rock-body may be recrystallized *in situ* into a coarse-grained rock—a plutonite—of essentially the same composition. Thus the recrystallization of the basal parts of a pile of basaltic lavas beneath a centre of eruption is a possibility, but there would be no petrogenetic significance attaching to this conversion: gabbro formed by recrystallization of basaltic lavas would not be significantly different from gabbro resulting from direct crystallization of basaltic magma.

With regard to norites, however, there is more to be said. The mineralogical differences between gabbro and norite may be correlated with chemical differences established by analysis; but it should be appreciated that the analyses available were made in the course of systematic studies of particular basic complexes. Gabbros grade into norites, and obviously there is little difference between a hypersthene-gabbro and an augite-norite: the *significant* differences

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would be apparent only by making a close comparison between orthogabbro and orthonorite. The chemical differences may indeed be anticipated from knowledge of the composition of augite and hypersthene. Norite should be richer in MgFe and correspondingly poor in CaAl, other things being equal. But the plagioclase, too, is involved. Plagioclase formed at high temperatures is anorthite-rich, *i.e.* it contains more than average CaAl, and therefore from magma thus depleted in these components hypersthene instead of augite should separate. Early crystallization of plagioclase of An-rich composition would only be possible from magma effectively super-saturated at high temperatures with these components, and H. H. Read believed that basaltic magma which had assimilated argillaceous sediment would give rise to norites. His special knowledge of these rocks resulted from studies of basic (gabbroic and noritic) intrusives occurring in eastern central Scotland. Among them are certain contaminated norites which yield clear evidence of magmatic assimilation of *argillaceous* materials. They contain very striking crystals of cordierite (illustrated in Fig. 53) associated with swarms of dark green octahedral spinels and red-brown mica additional to the basic plagioclase and pleochroic hypersthene proper to norite. Cordierite and spinel of this kind are metamorphic minerals, typical of thermally metamorphosed argillaceous sediments. They represent local and temporary excesses of Al resulting from the assimilation of xenoliths of nearby rocks of this composition. Therefore for the production of these particular norites the recipe was: local gabbroic (basaltic) magma + argillaceous xenoliths = cordierite-norite. It is significant in this connection that the greatest noritic complex in the world—the Bushveld lopolith—contains in its basal "envelope" rocks closely similar to the Scottish cordierite-norites in juxtaposition with high-grade sillimanite-hornfels representing aluminous sedimentary rocks below the base of the complex. The two occurrences (on very different scales) strengthen the validity of Read's conclusions; but two further comments may be made. The *time-factor* is important. In the early stages of formation of the Stillwater Complex the silicate phases precipitated were: first, olivine, followed by orthopyroxene (bronzite), followed in turn by augite. There is no time-scale available, but if there were, it would certainly show a wide separation in time between the first olivine crystals and the first precipitated augites. During that long period much MgFe was abstracted from the magma by the crystallization of olivine and bronzite, thus building up the requisite concentration of CaAl to permit the precipitation of augite. Thus the history of the Stillwater Complex commences with a chapter when the minerals appropriate to norite were precipitated. It will be realized, therefore, that norite can be produced

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from ordinary basaltic magma in the course of normal crystallization without the necessity of contamination by assimilation of argillaceous, or indeed any other extraneous material.

Troctolites are scarce in the greatest basic complexes—Bushveld, Stillwater and Sudbury, but they do fill an important role in others including notably the Freetown, Sierra Leone, and complexes in Rhum, Huntly and Belhelvie in Aberdeenshire. The interesting feature is the minor role played by pyroxenes which are absent from orthotroctolites. No troctolite represents the whole of the magma from which it was formed: no corresponding lava is known. But, as shown above, there is a time during the early stages of crystallization when, from an olivine-norite (or olivine-gabbro) magma, the only phases being precipitated may be olivine and CaAl-rich plagioclase, *i.e.* a troctolitic mineral assemblage.

Analyses of troctolites and other gabbroic and noritic rocks are tabulated, and these confirm the reality of chemical differences between these *rocks* that may be inferred from knowledge of their mineral composition. But these analyses (with one exception) tell us nothing of the composition of the *magma* from which they crystallized. One cannot infer the composition of a magma from that of a rock representing only a part of it; and therefore there is little point in comparing these analyses. Certain general facts may be stated, however. In a basic magma which is relatively deficient in SiO_2 the crystallization of olivine will be favoured. With sufficiently high SiO_2 an equilibrium assemblage will contain pyroxene and plagioclase but not olivine. Similarly, it is reasonable to infer that, other things being equal, to avoid the formation of clinopyroxene (augite) the magma must NOT contain more CaO and Al_2O_3 than can be accommodated in the anorthite-content of the plagioclase. With the passage of time it is known from physico-chemical considerations that CaAl decreases in the liquid phase as NaSi increases. Thus early crystal phases must consist of Ca-rich plagioclase with olivine (troctolitic assemblages); later come less calcic plagioclase (say bytownite) with bronzite or hypersthene (noritic assemblages); while relatively late phases will tend to consist of labradorite and augite (gabbroic assemblages).

How complete the sequence of rock-types is in any specific case will depend on (1) the composition of the magma; (2) the size of the magma body; (3) the temperature gradient (*i.e.* rate of loss of heat); and (4) whether or not there is any effective separation of solid phases and liquid. Obviously with the removal of liquid, reaction phenomena are curtailed and the rock assemblages will consist of early (high temperature) phases.

Gabbroic rocks in Britain additional to those mentioned above

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include the Crousa Down gabbro associated with a minor troctolite forming part of the Lizard Complex. Some of the larger basic minor intrusions of Ordovician age in Wales and the Welsh Borderlands are sufficiently coarse-grained to qualify as gabbros; but those which occur in the Tertiary ring-complexes in western Scotland and North-East Ireland are more noteworthy. They include the gabbroic complex which is magnificently exposed in the spectacular Cuillin Hills in Skye; the Ardnamurchan, Mull and Slieve Gullion Complexes which are described in Part IV in the final chapter.

Of uncertain age are the Carrock Fell gabbro in the Lake District (p. 447) and gabbros associated with diorites and granites in Guernsey and Jersey (Fig. 109). These latter rocks carry primary amphibole and would be termed *bojites* by some petrologists.

II. MICROGABBROS (DOLERITES)

The representatives of the gabbroic rocks which fall within the limits of the medium grain-size group should logically be termed microgabbros, as they are exactly comparable with microgranite, microsyenite and microdiorite. Unfortunately the logic of this was not realized sufficiently early, and two other names are in vogue at the present time. In Britain many petrologists use the name **dolerite** instead of microgabbro, though with different shades of meaning. To some the name implies *all* medium grained rocks of gabbroic composition, regardless of age, texture and mode of occurrence. Others restrict it to rocks of the appropriate composition exhibiting ophitic texture only; while yet others use it to cover such rocks provided they are of a hypabyssal mode of occurrence. In our view, if "dolerite" is to be used instead of "microgabbro," the name should cover *all* rocks of the appropriate composition irrespective of textural features, which can in any case be adequately covered by qualifiers. In America "diabase" replaces "dolerite"; but in this country, in spite of a veto by the Committee on Petrographic Nomenclature, some petrologists use the former term in the same sense as Rosenbusch did, for pre-Tertiary dolerites. That is, a **diabase** in this country is a rock of doleritic composition, which is altered to such an extent that few, if any, of the original minerals have survived. In view of these inconsistencies, there is a strong case for the general adoption of "microgabbro."

MINERAL COMPOSITION

So far as mineral contents are concerned, microgabbros (dolerites) closely resemble gabbros, norites, etc., and there is consequently no need to describe them in detail. The central type consists of plagioclase, near to labradorite in composition, clinopyroxene (usually common augite, though titanaugite often takes its place),

and iron-ore, which in different specimens may be magnetite, titanomagnetite, or ilmenite. The addition of olivine gives olivine-microgabbro (olivine-dolerite), while the incoming of quartz gives quartz-microgabbro (quartz-dolerite). The latter rocks are, of course, over-saturated, and in them orthopyroxene may replace, wholly or in part, the augite of typical microgabbro (dolerite). It is customary to lump together all dolerites containing hypersthene as "hypersthene-dolerites," regardless of the amount present. By comparison with their coarse-grained equivalents "hypersthene-dolerite" should indicate a rock of doleritic composition containing *accessory* hypersthene; but if the latter is of *essential* status, micronorite is the correct term to use. According to the degree of silica-saturation we may recognize quartz-micronorite, micronorite and olivine-micronorite. Ortho- and clino-pyroxenes may be present in all proportions, and as for the coarse-grained equivalents, the division is most conveniently drawn, quite arbitrarily, at 50 per cent. If therefore orthopyroxene is dominant, the rock is micronorite; but if clinopyroxene predominates, it is microgabbro. The qualifiers "mela-" and "leuco-" are used to signify respectively richness or deficiency in coloured minerals.

The occurrence of two or more pyroxenes in these rocks is a common feature. The clinopyroxene is frequently strongly schillerized parallel to (001) and when the crystal is in addition twinned on (100) it exhibits a distinctive herring-bone structure. The orthopyroxene is commonly hypersthene, appreciably, though faintly pleochroic in some sections, devoid of twinning unless it has inverted from pigeonite, and as a rule easily distinguished by its optical characters. Some crystals contain irregular cores of normal hypersthene which are surrounded by an irregular mantle of "hypersthene-perthite." This consists of hypersthene riddled with plates or irregular vermicules of clinopyroxene thrown out as a consequence of exsolution. All three types of pyroxene may be studied in sections of the Palisades sill, New York, which may be described as a quartz-hypersthene-microgabbro. The quartz in rocks of this type is particularly distinctive: it is intergrown graphically with orthoclase and occurs in angular interspaces between the lath-shaped labradorite crystals. In Britain this over-saturated type of micronorite is well represented by the famous Whin sill of northern England, and by the late-Carboniferous "quartz-dolerite" dykes in the Midland Valley of Scotland.

Usually hornblende does not occur in microgabbros; though a brown amphibole in some cases identified as barkevikite, but in others as "basaltic hornblende" (*i.e.* lamprobolite), does occasionally occur, as in the ~~minerite~~ *minerite* type, described from the parish of St. Minver in Cornwall.

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TEXTURAL VARIATION

As regards texture, microgabbros and micronorites are very variable. Without question the most characteristic texture is the *ophitic*, and some petrologists will not apply the name dolerite unless the rock is ophitic.

This texture, developed in a rock of the same composition but of coarse texture, is illustrated in Fig. 108. It will be noted that the plagioclases are euhedral, they are randomly arranged, and thus contrast with the clinopyroxenes which are relatively large and of irregular shape. Even in rocks of medium to fine grain individual pyroxenes may reach a diameter of as much as one inch and are therefore conspicuous in hand-specimens largely owing to the reflection of light from the cleavage surfaces.

If the plagioclase crystals penetrate into, but are not enclosed in, the pyroxenes, the texture is described as *subophitic*. Feldspars enclosed in the pyroxenes are frequently smaller than those outside, indicating that the periods of crystallization of the two minerals overlapped. This is probably the general case. Simultaneous crystallization produces this particular texture because of the inherent tendency for plagioclase to nucleate more readily than pyroxene and thus to set up many centres of crystallization, as compared with few, much more widely spaced centres of pyroxene-crystallization, the difference being of the order of twenty to one, as suggested by Hess, and indicated in Fig. 108. Less commonly the euhedral plagioclases may show a parallel orientation which must have been impressed at an early stage of crystallization as a result of flow movements in the melt. The orientation and size of the plagioclase crystals tend to be uniform both inside and outside the pyroxenes, suggesting that the latter developed at a somewhat later stage in the crystallization sequence.

The mineral composition affects the extent to which the ophitic texture is developed: whether plagioclase precedes or follows pyroxene is determined by the relative concentration of these two components in the magma. Plagioclase and pyroxene in a magma display a cotectic relationship as explained on p. 159, therefore whichever is in excess of the cotectic proportions will begin to crystallize first. In olivine-rich dolerites the crystallization of pyroxenes may be delayed relative to plagioclase with the result that the pyroxene is largely post-feldspar and the texture is ophitic.¹

Ophitic texture is not restricted to dolerites: some gabbros display this texture to perfection (Fig. 108), while in the fine-grain category certain basalts are "micro-ophitic."

¹ Walker, W. F., "Ophitic texture and basaltic crystallization," *Journ. Geol.* 65 (1957), pp. 1-14.

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Similarly, not all dolerites are ophitic. The term *intergranular* is used when the pyroxene forms grains interstitial to the plagioclase. This is frequently the case with dolerites of finer grain which grade into basalts and, like the latter, may be porphyritic. The term "labradorite-porphyry" is a legacy from the past and is best forgotten: "porphyritic dolerite" (or microgabbro) conveys all the necessary information in such cases.

Although some microgabbros (dolerites) of all ages are ideally fresh, many others are highly altered as a consequence of their subsequent treatment. Participation in earth movements, a mild degree of metamorphism and, naturally, weathering may lead to the replacement of any or all the original components, in the manner described above for the gabbroic rocks. Saussuritization, albitization, chloritization and epidotization may all contribute to the conversion of the original pyroxene, labradorite, iron-ore, etc., into albite, chlorite, epidote, calcite, leucoxene and quartz; but some of these minerals, though of late formation, are not secondary in the sense of having replaced pre-existing minerals. Thus although chlorite is widespread as an alteration product of pyroxene (with epidote, calcite, etc., as by-products), it occurs also in sharply defined interstitial areas between plagioclase and clinopyroxene which show no trace of alteration, and in these cases is primary. Rather special interest attaches to the quartz in these rocks. As noted above, it is a widespread primary constituent, normally intergrown with alkali-feldspar; but in addition it occurs also in irregular grains closely associated with other obviously secondary minerals, and is evidently an alteration product. Thirdly, quartz may be xenocrystic (see below, under "basalt"). Clearly it would be a mistake to refer to such rocks, with secondary or xenocrystic quartz, as quartz-microgabbros, or quartz-dolerites. Such terms should apply only to rocks containing the *primary* mineral.

Although amphiboles are rare primary constituents of the rocks under consideration, they are widespread as alteration products. Thus late-stage alteration tends to convert the original pyroxene into actinolite, an early stage showing perhaps merely a fringe of acicular crystals, but the alteration is progressive until all trace of pyroxene is lost, and a pseudomorph of closely packed fibres of pale-green amphibole ("uralite") is produced. In the process of dynamothermal metamorphism a compact common hornblende is produced from the pyroxene in the conversion of microgabbro into hornblende-schist or amphibolite. It follows that as a result of these changes the original labradorite-pyroxene combination gives place to common hornblende, a less basic plagioclase and various minor constituents to strike a balance. Thus the rock, although chemically gabbroic, in mineral contents is, in a broad sense,

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dioritic. To such rocks the term *epidiorite*¹ has been applied. Typical specimens occur among the sills in the South-West Highlands of Scotland.

It would be unprofitable at this stage to attempt a summary statement concerning the distribution of dolerites. Just as basalts are predominant in the extrusive, fine-grain category of igneous rocks, so dolerites are overwhelmingly dominant in the medium-grained rocks which normally are hypabyssal in mode of occurrence. This is emphasized by the immensely abundant dyke swarms and sill complexes which riddle the crust over vast areas. The most impressive example of the latter is afforded by the Karroo sills in South Africa, already mentioned, which underlie an area exceeding 10,000 square miles. An outstanding dyke swarm occurs in an east-west belt of country in Washington, U.S.A., some 50 miles by between 5 and 10 miles wide, which is occupied by thousands of vertical dykes of what is described as tholeiitic basalt, comparable in type with that which forms the flood basalts of the Columbia River area.² It consists of plagioclase forming half the rock, ranging in composition from An_{30} to An_{50} associated with both augite and pigeonite, together with interstitial glass.

The facts so far as Britain is concerned are included in the final part of this book; but passing reference may be made to the role of dolerites in the areas affected by Ordovician vulcanicity, notably in North Wales (the Dolgelley District) and the Welsh Borderland. The late-Carboniferous quartz-dolerite dykes with east-west trend are conspicuous in the Midland Valley of Scotland. Probably of the same general age is the only really extensive basic sill in the British Isles, that termed the Great Whin sill which underlies much of North-East England and is best known where its outcrop causes the High Force falls in Teesdale. The rock varies a good deal in the details of its mineral composition and texture; but it is essentially a quartz-hypersthene dolerite.³

Basalts are dominant in the Brito-Icelandic volcanic province, and dolerite dykes and sills might be expected. Actually the great majority of the hypabyssal rocks of the right composition are, in fact, basalts identical in composition and texture with the surface flows.

¹ "Epidiorite" (von Gumbel, 1874) is sometimes used in a wider sense, for example by Wiseman, J. D. H., who applies the term to all non-schistose rocks produced by dynamothermal metamorphism from basic igneous rocks, whether intrusive or extrusive. See "The Central and South-West Highland Epidiorites: A study in Progressive Metamorphism," *Quar. Journ. Geol. Soc.*, **90** (1934), p. 354.

² Foster, R. J., "The Teenaway dike swarm of Central Washington," *Amer. Journ. Sci.*, **256** (1958), pp. 644-53.

³ Holmes, A., and Harwood, H. F., "Age and Composition of the Whin Sill," *Min. Mag.*, **21** (1928), p. 493; and Tomkeieff, S. I., "Contribution to Petrology of Whin Sill," *Min. Mag.*, **22** (1929), p. 100.

III. BASALTS

The term "basalt" is one of the few rock-names familiar to the "man in the street," and is one of the oldest in petrology. It is applied collectively to the fine-grained equivalents of gabbros and norites. When fresh, basalts are black, minutely crystalline to compact rocks with high density. When exposed to weathering they assume a red or (more commonly) a greenish colour on account of the development of such secondary minerals as serpentines and chlorites. A further stage of alteration converts basalt ultimately into red bole.

MINERAL COMPOSITION

Microscopic examination of a typical basalt shows it to be composed essentially of plagioclase (generally within the range labradorite-bytownite), pyroxene and accessories, of which titanomagnetite is the most obvious. When olivine is present in addition, the rock is termed olivine-basalt.

As explained above (under "andesites") we are convinced that the only hope of achieving uniformity of nomenclature is to make use of clear-cut arbitrary definitions based on the composition of the plagioclases concerned. Thus in basalts the plagioclase *must* be more calcic than An_{50} . It is typically labradorite or bytownite, which may occur in two generations: as phenocrysts, often of relatively large size and commonly zoned, and as microlites of slightly more sodic composition.

Most basalts contain two kinds of pyroxene, the one kind being relatively calcic, and the other calcium-poor: the former is augite, and the latter may be pigeonite or an orthopyroxene depending upon the temperature of crystallization (see p. 46). Augite frequently occurs as phenocrysts which may show zoning and hour-glass structure; while second-generation crystals of minute size are commonly abundant in the groundmass. Pigeonite and orthopyroxenes are generally less conspicuous since they occur only as phenocrysts in basalts lying within quite a restricted range of composition. Pigeonite quite commonly occurs in the groundmass, though it is then extremely difficult to distinguish from augite.

One other factor tends to suppress pigeonite and orthopyroxenes: due to rapid chilling, olivine stands a better chance of survival than in a slowly cooled rock, in which the olivine may be converted by reaction into orthopyroxene. The survival of olivine in basalts must obviously affect the composition of the clinopyroxene. If magnesium silicate is locked up in the olivine, it is not available for the formation

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of pigeonite: in olivine-rich basalts, therefore, common augite tends to fill the role of pigeonite. Other things being equal, however, in an olivine-free basalt, the pyroxenes must be richer in magnesium, and pigeonite or orthopyroxenes become more important.

Hornblende is rare in basaltic rocks, but biotite in small quantities is not uncommon. Among the minor constituents iron ore is conspicuous, and is usually titano-magnetite in small octahedra; but dendritic iron ore often separates from the glassy base present in some basalts during the final stages of consolidation (Fig. 43C). Apatite is plentiful, though usually the crystals are minutely acicular. Secondary minerals are very varied. Olivine may show all stages of alteration to serpentine, talc, iddingsite, chlorophaeite, limonite or rhombohedral carbonate; pyroxenes are replaced progressively by chlorite with or without calcite and epidote; while the plagioclases undergo decomposition as described above for the gabbroic and microgabbroic rocks. Further, on account of the conditions under which they are erupted, basalts (even some dyke-basalts), tend to be vesicular, and although in recent specimens the vesicles are gas-filled, in the course of time they become filled with such minerals as chalcedony, agate, chlorite, calcite and especially zeolites including natrolite, phillipsite, heulandite and analcite (Fig. 57).

Occasionally isolated and much-corroded quartz grains occur in basalts which otherwise appear quite normal. These crystals are xenocrysts, caught up during the uprise of the magma. They are often surrounded by a reaction rim of sorts, consisting usually of closely packed granules of pyroxene. It is incorrect to call such rocks quartz-basalts: they should be termed **quartz-xenocryst-basalts**. Examples occur among the Permian lavas in Ayrshire, and in England among the Exeter lavas of the same age,¹ but such xenocrysts are liable to occur in any basalt, of any age.

Xenoliths (incorporated *rock-fragments* as distinct from crystal grains) are not uncommon and are of significance in that they provide valuable evidence of the nature of the rocks through which the magma passed in transit from its place of origin. Of special significance in this connection are the rock-fragments mistakenly recorded in the past as "olivine-nodules," but which are proved by examination of thin sections to be magmatically corroded pieces of the coarse-grained, ultrabasic rock, peridotite, consisting of the four minerals olivine, light greenish enstatite (practically indistinguishable from the olivine in hand-specimens), vivid green chrome-diopside and the accessory, chromite. The special significance of these peridotitic xenoliths is discussed below when dealing with magma-types.

¹ Cf. Tidmarsh, W. G., "The Permian Lavas of Devon," *Quar. Journ. Geol. Soc.*, **88** (1932), p. 741.

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TEXTURAL RANGE

In texture basalts are very variable: every gradation is represented between the vitreous—basalt-glass or *tachylyte*—and the holocrystalline. A large number of basalts contain glass forming minute angular patches between the crystals of the groundmass. Considerable interest attaches to the composition of this interstitial glass, and it is a feature that repays careful study with high magnification. Glass which may appear black and “dusty” may then be



FIG. 113

Variolitic basalt, near Rhobell Fawr, Merionethshire. Skeletal microphenocrysts of plagioclase set in variolitic groundmass of feldspar and augite. Small vesicles are present in the lower part of the field.

seen to contain innumerable minute octahedra of titano-magnetite, sometimes in parallel growth or arranged in a dendritic pattern. Equally minute granules of pyroxene are also found in some cases. These minerals indicate the strongly ferruginous character of some basalt glasses. Generally speaking iron-rich glass of this type constitutes only a very small proportion of the total rock, possibly because Fe tends to decrease the viscosity of silicate melts, and

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thus an abundance of Fe reduces the chance of a residual melt solidifying as glass. If there is an abundance of glass—and particularly if it has a low refractive index—there is a strong likelihood that the glass will be siliceous rather than ferruginous, because SiO_2 increases the viscosity of melts as witnessed by the glassy nature of so many acid volcanics. Naturally a siliceous and generally alkali-rich residuum is specially characteristic of the most siliceous basalts, the tholeiitic types described below. The character of the glassy mesostasis is of great significance in relation to differentiation of basaltic magmas and is discussed under that heading below.

Apart from forming these interstitial patches in otherwise crystalline basalts, tachylyte also forms the marginal parts of thin dykes and sills, and in the case of very thin sheets injected into cold rock, the whole may consist of tachylyte. Lava flows of basalt glass are rare, though not unknown, for example on Hawaii. The extraordinary material known as **Pelé's Hair** also comes from Hawaii. It consists of hair-like fibres of basalt glass, of an attractive golden brown colour, with occasional black swellings enclosing minute olivine crystals. It represents basaltic magma erupted as lava spray.

Variolitic texture is limited to basaltic rocks and is equivalent to spherulitic texture in the rhyolites. The essential feature is the occurrence of delicate brush- or fan-like "sprays" of radially disposed fibres of feldspar or less commonly of pyroxene (Fig. 113). Variolitic texture is especially characteristic of the crusts of pillow-lava. In some cases the varioles tend to stand out as small knobs on the weathered surface of flows of this kind.

Many basalts are porphyritic, with phenocrysts of any or all of the constituent minerals, generally identifiable in hand-specimens. They vary considerably in size and relative abundance, and different named types of basalt are distinguished by their phenocrysts, as explained below.

CLASSIFICATION OF BASALTS

A useful beginning in the classification of basalts involves a distinction between those that are over-saturated with respect to SiO_2 , and those that are under-saturated. Strictly a third category of just-saturated basalts should be recognized; but in practice, of course, this condition is very rarely achieved.

(1) **Over-saturated basalts** should theoretically contain no olivine, for there is sufficient silica to convert all of it into orthopyroxene. Quenching may prevent the reaction, however, and some olivine may survive. In this case surplus silica will be locked up in a residuum which, in fresh specimens, is glass. Separation and analysis of the latter have shown it to be of granitic composition, containing

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up to 70 per cent of silica.¹ These basalts are therefore the fine-grained equivalents of quartz-dolerites which, as we have seen, are characterized by the occurrence of interstitial mesostasis, consisting of alkali-feldspar and quartz, usually intergrown graphically. Because of the non-occurrence of olivine, the pyroxenes in over-saturated basalts are rich in MgFe—either pigeonite, or orthopyroxene or both, occurring as phenocrysts and small greenish prisms in the groundmass. Augite is generally less important and may be confined to the groundmass. Among over-saturated basalts **tholeiites**² hold the centre of the stage: in fact the latter term, after an obscure beginning, has come to be practically synonymous with “over-saturated basalts” and is of special significance in connection with certain petrogenetic problems considered later.

(2) **Under-saturated basalts** claim our attention next. They are easily identified on sight as they contain *abundant* olivine, though naturally the amount in a given specimen will depend on the degree of under-saturation. One other possibility must be noted—enrichment in olivine as a result of concentration of early-formed crystals (see picrite-basalt below).

The early separation of olivine in an olivine-basalt magma withdraws much of the available MgFe, so that when pyroxene is in turn precipitated, usually at a somewhat lower temperature, it will be a Ca-rich augite which separates, often as well formed phenocrysts. Groundmass pyroxene will also be augite typically, with or without pigeonite.³ In many olivine-basalts the clinopyroxene is titanaugite, rich in TiO_2 and Fe_2O_3 which may exsolve, forming the characteristic “fret” of regularly arranged minute inclusions.

Passing now from generic to specific characters, there is scope among under-saturated basalts for considerable variation, both in mineral composition and in texture. The former is by far the more important, and forms the basis of further classification. The three essential constituents vary in the roles they fill in different types: they may be grouped in different ways; and it happens that the chief groupings have been recognized among the basalts of Carboniferous age occurring in the Midland Valley of Scotland. These have been fully studied, adequately described and clearly defined. In our opinion there is no need to restrict the names applied to basalts in this one area and to lavas of this one age: they are recognizable

¹ Elliott, R. B., *Min. Mag.*, 31 (1956), p. 245.

² Jung, J. D., “Untersuchungen am Tholeiit von Tholey (Saar),” *Beiträge zur Min. und Pet.*, Bd. 6 (1958), pp. 147–81.

³ It would simplify the definitions if pigeonite were excluded; but we are influenced by a statement made by an authority on pyroxenes that he had never examined a basalt slide which does not contain *some* pigeonite. The position appears to be that over-saturated basalts contain *essential* pigeonite; but under-saturated types contain pigeonite only as a non-essential accessory.

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(with possibly minor variations) all over the world and there are many precedents for regarding these type-names as of general application. They are based on the kinds and proportions of phenocrysts present and four types are easily identified in the field. These are tabulated below:

Phenocrysts	Type
Olivine, augite, plagioclase (labradorite-bytownite)	Dunsapie (or Lion's Haunch)
Olivine and augite	Craiglockhart
Plagioclase only	Markle
Olivine and plagioclase	Jedburgh

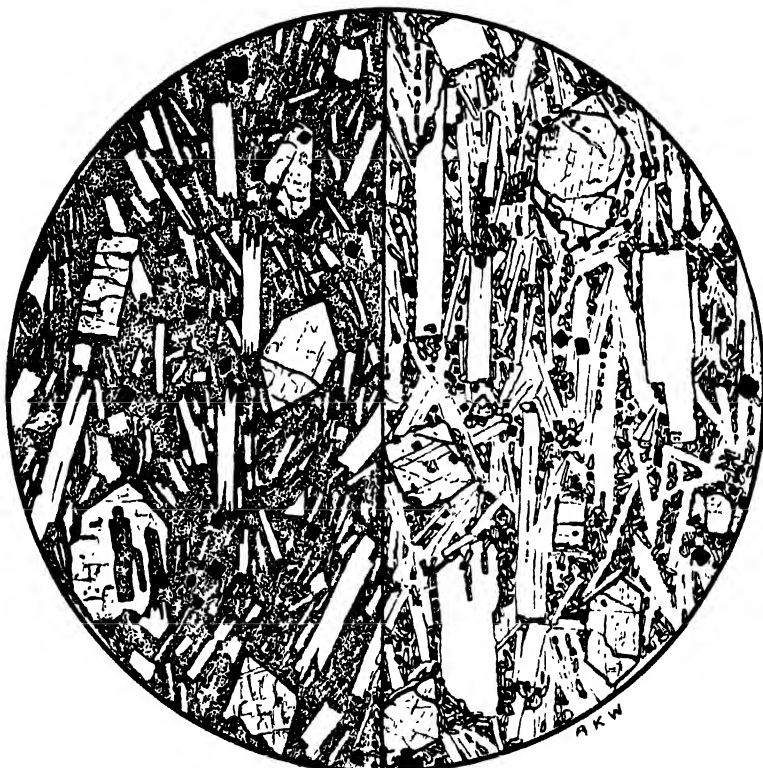


FIG. 114

Olivine-basalts from the Auvergne, France.

Left: Small phenocrysts of fresh olivine and plagioclase, many of the latter having forked ends. The groundmass is partially devitrified glass.

Right: Phenocrysts of olivine and plagioclase embedded in holocrystalline groundmass of lath-shaped plagioclase microlites, granules of augite and minute octahedra of magnetite. The texture is porphyritic, intergranular.

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In the original definitions the first three types were classified as **megaphyric**, containing large phenocrysts, while the fourth was a **microphyric** type. However a term is required for olivine-basalts containing olivine and plagioclase as the only phenocrysts present. In our view the size of the phenocrysts is of minor importance as compared with the facts of mineral composition. Two varieties of Jedburgh-type basalts from the Auvergne are illustrated in Fig. 114.

Melabasalts are so called as their content of mafic minerals is much higher than average. Two varieties are recognized: (1) picrite-basalt (also widely known as oceanite), olivine-rich; and (2) ankaramite, correspondingly rich in augite.

Picrite-basalt is best known from occurrences in Hawaii and other oceanic islands¹ (hence the name "oceanite"); but it is not restricted to them—picrite-basalt occurs also in such "continental" regions as the Deccan.² The quantity of olivine contained in picrite-basalt is far in excess of that occurring in a normal olivine-basalt, and may amount to about 50 per cent of the volume of the rock, with the olivines packed so tightly that they touch one another. There are good reasons for believing that picrite-basalt did not crystallize from a magma of its own composition, but resulted from the addition of olivine crystals to normal olivine-basalt magma, the olivines having been concentrated by gravity-controlled accumulation (see "Layering Phenomena" below). This is the mode of origin of picrites which, according to British usage, are coarse-grained rocks of essentially the same mineral composition, although the environments in which the two types originate are so different.

DISTRIBUTION OF BASALTS

Some of the oldest and most extensive eruptions of basalt are represented by the Keewatin and Keweenawan volcanics in the Pre-Cambrian of the Canadian Shield. Elsewhere, and during all geological periods basalts have been erupted over vast areas in many parts of the world, notably the Deccan (where they occupy approximately one-third of the total area of peninsular India, and locally reach nearly 6,000 feet in thickness), the Parana Basin in South America, the Snake River Plains in the U.S.A., and in the Thulean lava-field which includes Greenland, Iceland, the Faeroes and the coastal areas of North-East Ireland and western Scotland. The basalts in these areas form vast dissected plateaux, and are frequently referred to as flood- or plateau basalts. Elsewhere basalts

¹ Good descriptions of the general properties of the predominantly olivine-rich basalts of Hawaii are given by Macdonald, G.: "Hawaiian Petrographic Province," *Bull. Geol. Soc. Amer.*, 60 (1949), p. 1541.

² West, W. D., "The petrography of the . . . flows of Deccan Trap . . .," *Trans. Nat. Inst. Sci India.*, 4 (1958), pp. 1-56.

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have been erupted from cone-volcanoes of the central type, notably the Tertiary volcano of Mull in the Hebrides, and the active basaltic domes forming the Hawaiian Islands in the Pacific.

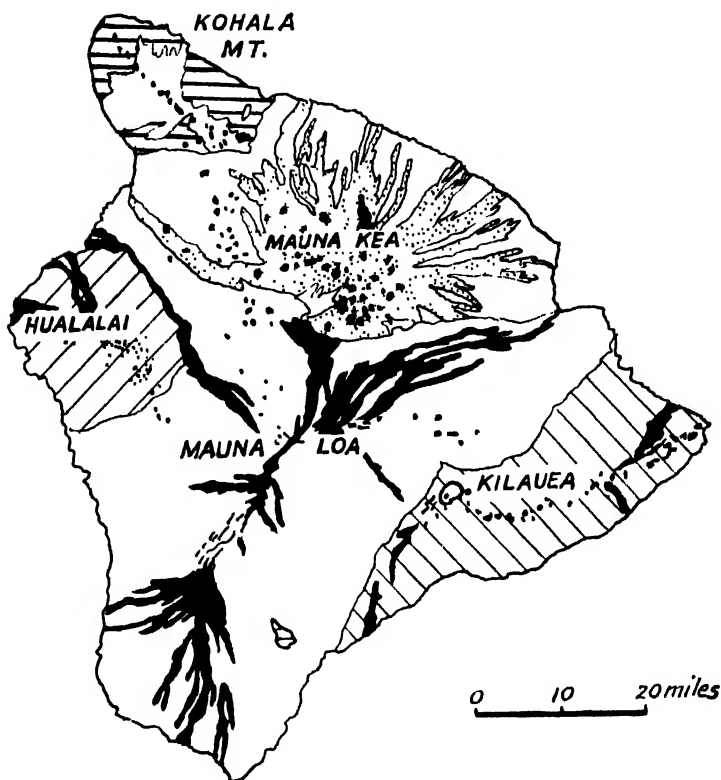


FIG. 115

Map of Hawaii showing distribution of lava flows, ranging in age from Pliocene (horizontal ruling) to Recent. Lavas belonging to the five main centres are shown by different ornament, while Recent flows are shown in black. Note the caldera of Kilauea within which is the Halemaumau Pit—a lava lake adjacent to which is the Volcano Observatory.

(After G. A. Macdonald.)

Rocks of identically the same compositional and textural range are widely distributed as dykes in these volcanic regions; they also form suites of cone-sheets, volcanic plugs and less regular vent-intrusions in volcanoes of central type.

In Britain basalts occur among the lavas of all the volcanic episodes from the Ordovician to the Tertiary, but are particularly

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well developed among the lavas of Carboniferous (including Permo-Carboniferous) and Tertiary age. The former are well exposed and widely distributed in the Midland Valley of Scotland, particularly striking features including the scarp-terracing of the basaltic Campsie Fells, and the isolated "volcano" of Arthur's Seat which dominates the scenery of the Edinburgh district (Figs. 54, 55).

The Tertiary basalts of this country have been studied in great detail, and even the "man in the street" is familiar with their general nature and significance largely as a result of the spectacular columnar jointing so magnificently displayed at the Giant's Causeway in Co. Antrim (Fig. 58) and on the island of Staffa (Fingal's Cave) in the Hebrides. The important part played by the Tertiary basalts in the development of ideas concerning magma-types is explained below.

Summary accounts of these and other basalts are included in the appropriate chapters in the last part of this book.

ORIGIN AND DIFFERENTIATION OF
BASALTIC MAGMAS

IN view of their widespread distribution outlined in the previous chapter, basalts have generally been considered to represent a largely undifferentiated type of magma which has been readily available throughout geological time. Daly regarded the source of basaltic magmas as being the sima, an enveloping earth shell underlying the ocean basins, and in the continental regions lying beneath the sial. He further suggested that the sima might have the properties of basaltic glass, because a glass is easily converted into a liquid of its own composition either by rising temperature or by the combined effects of temperature and pressure variations. Modern geophysical evidence, however, suggests that the sima and the sial (*i.e.* the crust of the earth) are entirely crystalline. To convert crystalline rock into magma involves a more complex process with, initially, selective fusion of the components which melt at the lowest temperatures. Ultimately, of course, all the original rock components may become molten; but until that point is reached there may be abundant opportunity for liquid fractions of varying composition to become separated from the still solid material. This could well be a cause of some variation in the composition of basalts. Another important factor is the undoubted vertical and lateral variation in the composition of the crustal rocks from which basalt magmas may be derived. In recent years increasing attention has been paid to a possibility, first proposed by Bowen, that basaltic magma may be derived by selective fusion of the upper layers of the mantle, the olivine-rich and ultrabasic earth shell underlying the crust. Thus, although the range of composition of basalts is very limited, it is not surprising that minor differences are displayed by basalts belonging to any given petrographic province and between one province and another.

Many eminent geologists, among them Daly¹ and Bowen² have deduced that virtually the whole range of igneous rock-types could be derived, either by differentiation or syntexis, from one kind of

¹ "Igneous Rocks and the Depths of the Earth" (1933), pp. 189, 304, 395.

² "The Evolution of the Igneous Rocks" (1928), pp. 5, 21, 320.

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magma, essentially of basaltic composition. Subsequently this concept of the parental status of basaltic magmas has been elaborated, mainly towards the recognition of a number of chemically distinctive basaltic **magma-types**.

The first use of this term was made by the authors of the Mull Memoir, who distinguished three basaltic magma-types which they regarded as parental to the lavas erupted in Tertiary times from the Mull volcano.¹

The three types were named: (1) Plateau-Basalt Magma Type; (2) Central Non-Porphyrific Magma Type; (3) Central Porphyritic Magma Type. The choice of names was unfortunate in several respects,² notably so in the case of the "Plateau Basalt" type. Although it is convenient to apply the term plateau basalt collectively to basalts which build the great lava plateaux of the world, it can strictly be used only in a topographic, not petrographic, sense: for this reason "flood basalt" is a better general term. So far as is known *any* variety of basalt may occur as plateau basalt, although probably on average over-saturated tholeiitic basalts predominate (e.g. in Oregon). It can only lead to confusion, therefore, to use the term plateau basalt petrographically. In the case of Mull the anomaly is increased by the fact that the plateau basalts there are predominantly under-saturated olivine-basalts and are therefore atypical.

The possibility of linking the various dominant suites of associated lavas with parental magmas from which they were derived was explored by Kennedy who concluded from the facts of distribution that two such magma-types were involved.³ These were identified with two of the Mull magma-types as follows: Olivine-Basalt Magma type (equivalent to the so-called Plateau type of Mull), and Tholeiitic Magma type (equivalent to the Central Non-Porphyrific type).

The essential chemical characteristics of the two basaltic magma-types recognized by Kennedy are tabulated opposite.

Olivine-basalts are particularly abundant in the volcanic islands of the ocean basins, notably in Hawaii, where they have been most completely studied. The dominant olivine-basalts are associated with less basic lavas believed to be differentiates of the parental magma. This suite has come to be known as the Oceanic Suite, but this term is somewhat of a misnomer since similar lavas are equally well represented in the stable continental regions: they are therefore

¹ Bailey, E. B. and others, "The Tertiary and Post-Tertiary Geology of Mull," *Mem. Geol. Surv. Scotland* (1924).

² See Wells, M. K., and Wells, A. K., "Magma-types and their Nomenclature," *Geol. Mag.*, 85 (1948), pp. 349-57.

³ See Kennedy, W. Q., and Anderson, E. M., "Crustal Layers and the Origin of Magmas," *Bull. Volc. Série ii, Tome ii* (1938), p. 24.

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world-wide or "universal" in the popular sense. The rocks involved typically comprise: (alkali) olivine-basalt, hawaiiite, mugearite and sodic trachyte. Less frequently feldspathoidal lavas—especially phonolites—also occur among the lavas of the oceanic islands, although in some continental regions, particularly where there is rift faulting, feldspathoidal types may be very abundant.

	Olivine-Basalt Magma-Type.	Tholeiitic Basalt Magma-Type.
SiO ₂	45	50
Al ₂ O ₃	15	13
FeO+Fe ₂ O ₃ . .	13	13
MgO	8	5
CaO	9	10
Na ₂ O	2.5	2.8
K ₂ O	0.5	1.2

Melabasalts of both olivine-rich and augite-rich varieties are distinctive members of the suite. The corresponding suite which is restricted to continental regions comprises basalt, andesite (true), dacite and rhyolite: these may appropriately be regarded as the typical members of the Continental Suite. Two other terms are in common use and were introduced, presumably, to indicate the essential chemical characteristics of the two suites. They are "Alkali Suite" and "Calc-Alkali Suite," the latter being applied to the Continental, and the former to the so-called Oceanic Suite.¹

The significant facts regarding the distribution of andesites and of rhyolites have been discussed in previous chapters: at this juncture we are concerned with the relationship of these rocks to basalts, and in particular with Kennedy's suggestion that the members of the Continental Suite might be related to a parental tholeiitic magma. This hypothesis is based upon a number of facts of which the following are perhaps the most significant: tholeiitic basalts are particularly abundant in continental regions, constituting the dominant types of flood basalts; while quartz-dolerites of identical composition occur in innumerable minor intrusions in many associated

¹ Especially to readers of this textbook these two terms are unfortunate, as, used in this sense, they convey a wrong impression of the chemical characteristics of some of the rock-types covered by the terms. It is absurd to apply the term "alkali" or "alkaline" to ordinary olivine-basalt and melabasalt; and just as unreasonable to label rhyolite as "calc-alkaline"; but it is the current practice. Because they belong to the "Alkali Suite" the typical olivine-basalts of Hawaii and elsewhere are now distinguished as "alkali olivine-basalts," although such rocks are among the most typically calc-alkaline in composition.

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dyke swarms and in the sills of the Karroo, South Africa; there is often a very close field association between tholeiites and rocks of rhyolitic composition, for example in composite dykes and sills; fractional crystallization of a tholeiite leaves a residual liquid which congeals to interstitial glass of rhyolitic composition, as explained in the previous chapter. Despite this apparently convincing evidence, amongst the vast accumulation of more recent data there are new facts concerning the composition and distribution of lavas which make it appear unlikely that the rock-types comprising the Continental Suite are derived, by differentiation, from a tholeiitic parental magma.

According to the two magma-type hypotheses, olivine-basalt magma is derived from a continuous earth shell underlying the oceans and the continents; while tholeiitic magma is derived from an overlying layer, sandwiched between the olivine-basalt layer and the granitic sial, and like the latter confined to continental regions. It is reasonable to suppose that the relatively siliceous tholeiites may owe their origin to syntexis involving olivine-basalt magma and sialic rocks. However, there is now some evidence that rocks having tholeiitic affinities occur among the early eruptives of Hawaii and in other oceanic islands¹ where presumably the sial is missing. Evidence on this point is difficult to assess since it rests in part on the degree of affinity of the rocks concerned with genuine continental tholeiites: questions of definition are involved. In this connection it must always be remembered that knowledge of oceanic volcanic rocks is severely restricted, being confined to the unsubmerged tops of the largest volcanoes. However, the evidence is strong enough for Macdonald² to list tholeiitic basalts as members of his "oceanic suite," together with the dominant olivine-basalts and their derivatives already listed. It is significant that even if tholeiite basalts do occur in the ocean basins, true andesites, dacites and rhyolites do *not* occur with them. In other words, it appears likely from the evidence of the oceanic volcanic rocks that differentiation of basaltic magma (particularly the tholeiitic type) does not produce andesites unless syntexis with sialic material is also operative. This is in agreement with conclusions reached on chemical grounds, as discussed in the account of andesites; and it finds further support from the distribution and association of tholeiites with other volcanic rocks in continental environments.

A clear-cut distinction must be drawn between the volcanic rocks of stable and orogenic continental regions. It is generally only in

¹ Tilley, C. E., "Some aspects of magmatic evolution," *Quar. Journ. Geol. Soc.*, **106** (1950), pp. 37-61.

² Macdonald, G. A., "Dissimilarity between continental and oceanic rock-types," *Journ. Petr.*, **1** (1960), pp. 172-7.

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the former that flood basalts and their associated intrusive rocks are found, these often, though by no means exclusively, being of tholeiitic type. Intermediate volcanic rocks are generally present in subordinate amounts only, and further exhibit mineralogical features, such as the occurrence of potassic feldspar, which distinguish them from normal orogenic andesites. Evidence relating to the course of differentiation during fractional crystallization of basic magmas in general in relatively stable continental regions is provided by the sequence of rock types occurring in some of the layered complexes discussed in a later chapter. In a number of cases there is a relatively sharp transition between the gabbroic or noritic rocks forming the lower layers and the granites or granophyres which may form a capping, so that rocks of "intermediate" composition are virtually excluded, just as they are from the corresponding lava successions.¹ Iron enrichment is a characteristic feature of the successive differentiates in layered complexes, particularly in the basic members; while the chemical variation towards the acid end of the series is generally quite different from that of the andesite-dacite-rhyolite series. In non-orogenic regions, therefore, rocks of tholeiitic composition may predominate, and although acid differentiates commonly accompany them, "intermediate" rocks tend to be suppressed. There is certainly no evidence that differentiation of a tholeiitic magma-type gives rise to andesites.

It is unnecessary to add much about the orogenic volcanic suites of the continental regions in view of the accounts already given in earlier chapters of andesites and rhyolites. Within this type of environment—which includes many of the island arcs—andesites are frequently more abundant than basalts. The basalts which occur in this association are variable in character and show affinity with tholeiites only occasionally. Many of them are porphyritic and suggest an accumulative origin for the phenocrystic constituents, which may include a high percentage of plagioclase.² It is largely the latter which accounts for the high content of alumina in many of these basalts. It has recently been suggested that a third distinctive magma-type should be recognized, to include these high-alumina basalts and their derivatives.³ In Japan they occupy a zone between the oceanic and the continental regions, as well as being intermediate in character between the two usually recognized suites. This discovery still further weakens the two-basaltic magma-types hypothesis. It may well be that too much emphasis has been placed upon fundamental differences between magma-types: as more analytical data become available it appears that

¹ Hamilton, W., "Silicic differentiates of lololiths," *Int. Geol. Cong.*, xxi Session, Pt. 13 (1960), pp. 59–67.

² Tilley, C. E., *op. cit.*, p. 54.

³ Kuno, H., "High-alumina basalt," *Journ. Petr.*, 1 (1960), pp. 121–45.

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basaltic compositions cover a continuous spectrum. On average the basalts from any one region tend to have some degree of distinctiveness when compared with those from another. Such regional differences as the undersaturated condition of the olivine-basalts

CHEMICAL ANALYSES OF TYPICAL BASALTS

	I. Tholeiite.	II. High-alumina basalt.	III. Picrite-basalt.	IV. Olivine- basalt, Hawaii.
SiO ₂	49.78	50.19	46.62	48.35
Al ₂ O ₃	15.69	17.58	8.68	13.18
Fe ₂ O ₃	2.73	2.84	2.04	2.35
FeO	9.20	7.19	10.52	9.08
MgO	7.79	7.39	20.86	9.72
CaO	11.93	10.50	7.15	10.34
Na ₂ O	1.21	2.75	1.41	2.42
K ₂ O	0.29	0.40	0.28	0.58
H ₂ O	—	—	0.23	—
TiO ₂	0.68	0.75	1.71	2.77
P ₂ O ₅	0.07	0.14	0.14	0.34
MnO	0.35	0.25	0.14	0.14
Other constituents .	0.29	—	0.22	0.73
	100.01	99.98	100.00	100.00

- I. Average of 3 tholeiites, Japan and Korea, quoted from H. Kuno, *Journ. Petr.*, 1 (1960), p. 141.
- II. Average of 11 high-alumina basalts, Japan and Korea, quoted from H. Kuno, *loc. cit.*
- III. Picrite-basalt, average of three analyses, Hawaii, quoted from R. A. Daly (1933), p. 397.
- IV. Average of 53 olivine-basalts, Hawaii, quoted from G. A. Macdonald, *Bull. Geol. Soc. Amer.*, 60 (1949).

of the oceanic islands as compared with the oversaturated tholeiitic basalts of non-orogenic continental regions does not *prove* the existence of different magma-types: the differences, we believe, may have arisen as a consequence of different processes to which basaltic magma in general may be subjected before it becomes available at the surface for examination.

Finally reference may be made to the outstanding difficulty encountered in studying magma-types. They cannot be isolated and studied chemically. A magma-type is an abstract conception, which cannot, therefore, be analysed. The best that can be done is to concentrate on basalts as being nearest in composition to the parental magmas. Analyses purporting to represent the compositions

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of the various magma-types—the term parental magmas is perhaps more appropriate in this connection—are available; but they are merely averages of a number of selected basalts of a given type.

What relationship these averages bear to the original magmas cannot be known: it probably depends upon how far back in time one pursues the problem. In petrogenitic studies basaltic magmas do not represent the real beginning, but their availability marks the end of the first stage in the evolution of igneous rocks. Where and how such magmas are produced are elusive problems which cannot be studied by direct observation. Suggested solutions are therefore necessarily highly speculative and lie within the province of the geophysicist rather than the petrologist.

SPILITES AND ALLIED ROCK-TYPES

THE status and significance of spilites are matters of dispute among petrologists, while their position in a scheme of classification is problematical.¹

Spilites (Brogniart, 1827) are basic, fine-grained rocks, with SiO_2 percentage averaging 40, occurring as pillow-lavas of various ages in many parts of the world. Petrographically they are distinctive as the feldspar they contain is albite, in spite of their basicity. Secondly, the place of the usual ferromagnesian silicates is taken by chlorite, often of an iron-rich type, though augite does occasionally occur.

Intermediate and acid lavas and minor intrusions are associated with spilites in the way that basalts are accompanied by subordinate trachytes and rhyolites. These are known as keratophyres and quartz-keratophyres respectively.

Further, in areas where spilites, keratophyres and quartz-keratophyres occur as lavas, certain sills and dykes of spilite occur, passing into albite-dolerites of coarser grain, while some are of sufficiently coarse grain as to fall into the albite-gabbro category. A variant of albite-dolerite is minverite described above. Other members of the suite, of ultra-mafic composition, include augite-picrite at Menheniot, Cornwall, and hornblende-augite-picrite at Molenick in the Plymouth district. Thus there is a group of rocks proved by their distribution to be comagmatic, linked to one another by distinctive chemical and mineralogical features. These rocks constitute the Spilitic Suite of H. Dewey and Sir John Flett who emphasized the strongly sodic character of the several members of the suite.²

With regard to classification the whole range is alkaline with special emphasis on high soda-content and notable deficiency in

¹ Wells, A. K., "The Problem of the Spilites," *Geol. Mag.* (1923), p. 63; and Sundius, N., "On the Spilitic Rocks," *Geol. Mag.* (1930), p. 1; Vuagnat, M., "Sur quelques diabases suisses: contribution à l'étude du problème des spilites et des pillow lavas," *Bull. Suisse de Min. et Pétrol.*, 26 (1946), p. 155. Bailey, E. B., and McCallien, W. J., "Some aspects of the Steinmann Trinity, mainly chemical," *Quar. Jour. Geol. Soc.*, 116 (1960), pp. 365-95. The reader is recommended to study Turner and Verhoogen's chapter in *Igneous and Metamorphic Petrology* (1960), p. 258.

² "Some British pillow lavas . . .," *Geol. Mag.* (1911), pp. 202-9 and 241-8.

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potash. A point we must emphasize is one that is liable to be overlooked: the terms keratophyre and quartz-keratophyre are not synonymous with sodic Intermediate lavas (soda-trachytes) and sodic Acid lavas (soda-rhyolites) respectively. Among Na-trachytes, for example, one type is appropriately termed keratophyre; but there are other types which are definitely *not* keratophyres. To be entitled, so to speak, to the name "keratophyre," an albite-trachyte must be associated with other members of the Spilitic Suite in the right kind of environment.

There are considerable difficulties in the way of stating the true facts regarding distribution, since many petrologists refuse to recognize spilite as a valid type, and claim that these rocks are merely "altered basalts." It may well be, therefore, that spilites and other members of the suite are more widely distributed than available records would suggest. The distinctive chemical features, especially the high sodium content, are claimed to have been impressed upon magma developing in a geosynclinal environment and subsequently erupted, perhaps from fissures opening in the sea-floor. Certainly spilites are interbedded with sediments of geosynclinal types, while the frequency with which they have developed pillow structure proves them to have been of submarine origin. The material filling the interstices between the pillows is dark blue-black mudstone in some occurrences, but chert in others. In some instances the chert is a banded silica-haematite rock of jaspilite type. Occasionally jaspilite occurs *within* the lava pillows proving that it was caught up and incorporated at the time of eruption. The most extensive occurrences of spilitic rocks are found in New South Wales and New Zealand and have been thoroughly studied by W. N. Benson and others. In Britain spilites have been described from among the Pre-Cambrian, Ordovician and Devonian-Carboniferous volcanic assemblages. Spilites are well developed on several horizons in the Mona Complex of Pre-Cambrian age in Anglesey.¹ A division of the Dalradian at Tayvallich in Argyllshire includes spilitic lava flows and intrusive basic sills of epidiorite. Notable occurrences are those of Ordovician age to be found on the Girvan coast in south-western Scotland; those of the same period in parts of North Wales, notably on Cader Idris in the Dolgelley area, and of Carboniferous age in parts of Devon (Chibley) and Cornwall. In the latter county a spectacular display occurs at Pentire Point near Padstow.

With regard to origin it is now more generally agreed than formerly that spilites and the Spilitic Suite do really exist, though opinion is divided as to their mode of origin. The main question at issue is

¹ Striking photographs as well as a detailed account of the field relations and petrography will be found in Greenly, E., *Memoir on Anglesey*, Vol. 1 (1919), pp. 54 and 71.

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whether their distinctive compositional character is a magmatic quality or one impressed later as a consequence of metasomatism involving albitization.

Spilites *are* restricted to geosynclinal regions; they were erupted under submarine conditions. They are markedly vesicular, and although the vesicle-infilling materials include various forms of silica (chalcedonite and quartz), chlorite and calcite, the last-named is the most abundant and accounts largely for the relatively high CO_2 (10 per cent) shown on analysis. It cannot be ascertained whether these substances were derived from internal or external sources; but certain spilites in Anglesey have been extensively carbonated, in some cases to such an extent that only a shell of spilite remains. The rest is chiefly carbonate. Similarly some spilites in the Dolgelley district are completely pseudomorphed by carbonates though the characteristic texture has survived. Therefore in the latter cases the replacement of the original minerals by calcite was a secondary, post-crystallization process.

Some of the calcite, epidote, etc., in spilitic rocks may have resulted from *internal rearrangement* of components. The most significant possibility concerns the albite and calcite present in the rocks. If the Ca were available for building feldspar instead of being immobilized in the calcite, etc., it would allow crystallization of a more normal basaltic type of plagioclase.

Many of the petrographic features of spilites are comparable with those seen in basalts that have been subjected to low grade regional metamorphism. Although relics of primary augite and calcic plagioclase may be present they are often entirely absent. In the latter case there may be no evidence to show whether the albite is of secondary origin or has crystallized directly from the magma.

Much of the albite, calcite and epidote may have resulted from a purely internal rearrangement of the components of calcic plagioclase, so that the chemical composition of a spilite may not be greatly different from that of basalt. Spilitic magma, by inference, may be regarded as essentially basaltic, but with an abnormally high H_2O —and probably CO_2 —content. Crystallization of such a magma would follow a course different from that appropriate to normal basalt, particularly if the escape of volatiles were prevented either by the rapid formation of an impermeable envelope of glass in pillow-lava, or by some other factor associated with a submarine environment.

It may be noted that P. Eskola and others¹ successfully brought about albitization of calcic plagioclase by the action of Na_2CO_3 solutions, which proved that albitization such as that seen in

¹ Eskola, P., Vuoristo, U., and Rankama, K., "An experimental illustration of the spilite reaction," *Comptes rendus Soc. Géol. Finlande*, No. 9 (1935), p. 1.

THE SPILITES AND ALLIED ROCK-TYPES

basalts may be effected at temperatures as low as 250° C. This is a suggestive experiment, though it does not prove that all, or indeed that any, spilites were actually formed in this way.

It is very significant that shales marginal to spilitic minor intrusions (of albite-dolerite, etc.) may be converted into **adinole**. This is a metasomatic, compact white-weathering rock consisting of up to 90 per cent or more of albite. Adinoles provide the field evidence for the existence of sodic solutions emanating from the spilite magma and capable of metasomatizing the country-rock. Similar mineralogical changes brought about by the solutions trapped within the spilites may be regarded as auto-metasomatic.

It appears, therefore, that, as a consequence of down-warping in the early stages of the development of a geosyncline, basaltic magma is locally generated by selective fusion of a peridotitic earth-shell. During uprise through unindurated sediments containing juvenile water this magma developed its hydrous and essentially spilitic quality; and crystallization under the special environmental conditions obtaining during the submarine eruption of pillow-lavas accounts for the distinctive mineralogical features of spilites. It has been suggested that under hydrous conditions metastable augite would crystallize together with albite, followed by primary Fe-rich chlorite and quartz. Retention of volatiles is essential, and neither introduction of Na from some extraneous source nor the existence of a special kind of magma appears necessary.

SYENOGABBROS AND TRACHYBASALTS

1. SYENOGABBROS

AMONG both the basic plutonites and lavas of the same composition rocks occur in which the characters of syenites and gabbros on the one hand, and trachytes and basalts on the other, are combined. Within the general basic framework, affinity with the former should be shown by the occurrence of plagioclase near labradorite in composition. The fundamental facts, then, are two in number: firstly these rocks are thoroughly basic as shown chiefly by their high colour index. Secondly, alkali- and calc-alkali feldspar occur in association, and ideally in nearly equal amounts. These facts suggest that the best group name available is syenogabbro for the coarse-grained, and trachybasalt for the fine-grained, members.

The alkali-rich portion of any member of this series may, in theory, consist of an alkali feldspar, or its unsaturated equivalent leucite (in fine-grained members of the series) or nepheline. The gabbroic (or basaltic) portion consists of plagioclase plus mafics, of which pyroxene is the commonest.

Saturated varieties are by no means common rocks in the coarse-grain category; but the Scottish type kentallenite is typical syenogabbro and a very distinctive rock.

Kentallenite¹ was discovered at Kentallen near Ballachulish in western Scotland. It is a heavy, dark-coloured rock, with prominent patches of bronzy mica and an abundance of black crystals embedded in a rather meagre feldspathic base. In thin section olivine is very prominent, chiefly as it is densely charged with a separation of magnetite, as a fine dust that imparts a general grey colour to the crystals, and also as irregular dendritic patches. The most abundant mineral is a light greenish augite, in euhedral crystals. The feldspar is not easy to deal with; but careful examination will usually show that both plagioclase and orthoclase are present, involved in the same sort of relationship as in monzonite, but rather less obviously. All of these features are illustrated in Fig. 116. The

¹ Hill, J. B., and Kynaston, H., "On Kentallenite and its Relation to other Rocks," *Quar. Journ. Geol. Soc.*, 56 (1900), p. 531.



FIG. 116

Kentallenite, the type-rock from Kentallen, Argyllshire, Scotland.

Olivine (close stipple) enclosing dendritic magnetite; augite (light stipple); biotite (lined); plagioclase shown in outline with traces of the twinning orthoclase left blank; apatite also shown.

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relationships between the mafic minerals are interesting and unusual. Olivine crystals are not notably euhedral: they are sometimes bounded by crystal-faces, but in the main give the impression of being somewhat rounded and "corroded." By contrast the greenish augites are singularly well-formed. The abundant mica is obviously related to the olivines which in many cases are partially or completely enveloped in the former, which is poikilitic and highly irregular in shape. It is clear from the relationship between the feldspar and the biotite that the two crystallized together. The biotite was obviously late in the crystallization sequence.

Normally in a crystallizing basic magma the mafic minerals follow one another in the order of Bowen's Discontinuous Reaction Series. In kentallenite, however, the sequence is broken by a wide gap in the middle of the sequence: orthopyroxene, clinopyroxene (in its normal reaction relationship with olivine), and amphibole are unrepresented: on the other hand the last member of the sequence (biotite) is seen in close juxtaposition with the first—the olivine. These are significant features: the reader is invited to draw his own conclusions, preferably after a careful examination of the evidence afforded by thin sections.

It is extremely unlikely that all rocks to which the term syenogabbro might be applied conform to the kentallenite pattern. Kentallenite is a very distinctive rock, but all syenogabbros are not kentallenites. The rarity of these rocks suggests that they are brought into being by an unusual combination of circumstances.

Syenogabbros which are undersaturated in respect of their felsic constituents include gabbroic rocks containing nepheline and/or analcite. The types essexite, theralite, teschenite, crinanite and lugarite all fall in this category.

Essexite is one of the least satisfactorily defined of the basic rocks, and the name is applied to widely different mineral assemblages. The original essexite occurs in association with nepheline-syenites at Salem Neck,¹ Essex County, Massachusetts, and was defined in rather general terms by J. H. Sears. On re-examination the type-rock proved to have suffered contact metamorphism and metasomatism by an adjacent intrusion. Fundamentally essexite is a dark-coloured gabbroic rock in which a varied assortment of coloured silicates may accompany plagioclase, typically near labradorite, and small amounts of alkali-feldspar and feldspathoids. With regard to the last named, opinion seems to have been far from unanimous as to the amount of nepheline and/or analcite which

¹ Washington, H. S., "The Petrographical Province of Essex County, Mass.," *Journ. Geol.*, 7 (1899), p. 53.

SYENOGABBROS, MICROSYENOGABBROS, TRACHYBASALTS

should be present. Certain so-called essexites from Norway, described by Brøgger, are not notably different from rocks which others would term olivine-gabbros: they contain no nepheline. Another well-known essexite from Rongstock in Bohemia contains only trifling amounts of that mineral; but in the Scottish essexites nepheline can be discovered in thin sections, although it has to be searched for. But although the mere list of minerals present does not suggest a highly distinctive rock-type, actually the Scottish essexites are well characterized, though it is textural quality rather than mineral content that is important (Fig. 117). In so far as the rocks consist chiefly of labradorite, titanaugite, olivine, apatite and ilmenite, they are like many another olivine-gabbro; but tucked away in the interstices between the laths of labradorite are patches of nepheline and of unmistakable analcite. Further, the olivines are fresh and clear, the titanaugites particularly well-formed, and at once attract attention in the hand-specimen. These strongly porphyritic essexites are well represented by the Crawfordjohn rock well-known to many Scotsmen, as it is a favourite "curling-stone."¹

By increase in the amount of nepheline essexite passes into theralite, described below.

Teschenite (Hohenegger, 1861) is so named after the original locality, Teschen, in Silesia. In mineral composition teschenite is in general like gabbro, but differs in containing analcite as an essential constituent. The type is therefore defined as consisting essentially of basic plagioclase, near labradorite, clinopyroxene (usually titanaugite), and analcite, together with the usual gabbroic accessory minerals. Commonly barkevikitic amphibole occurs in close association with the titanaugite, and biotite often occurs in small quantity.² Olivine is not constantly present, and is therefore omitted from the definition of the type; when olivine does occur, the rock is **olivine-teschenite**.

In texture teschenites vary widely: they may be ultra-coarse—teschenite-pegmatites, coarse to medium—the majority hover on the brink between gabbroid and doleritic, while some are notably finer than the main body of the intrusion, of which they form a part.

In Scotland teschenitic rocks form large, often differentiated sills, such as the Inchcolm sill³ which forms the island of that name in the Firth of Forth (see map, Fig. 140), and the well-known sheet exposed at Salisbury Crags, Edinburgh (Fig. 54). Certain of the

¹ Scott, A., "The Crawfordjohn Essexite and Associated Rocks," *Geol. Mag.*, (1915), p. 458.

² Tyrrell, G. W., "Classification and Age of the Analcite-bearing Rocks of Scotland," *Geol. Mag.*, 60 (1923), pp. 249–60; F. Walker, "Notes on the Scottish and Moravian Teschenites," *Geol. Mag.*, 60 (1923), pp. 242–9.

³ Campbell, R., and Stenhouse, A. G., *Trans. Geol. Soc. Edin.*, 9 (1907), p. 121.



FIG. 117

Essexite, Crawfordjohn, Scotland.

Euhedral olivines and clinopyroxenes set in a groundmass of labradorite, nepheline (turbid through alteration), clear interstitial analcite, magnetite, and apatite, in some instances, cored. Note the zonal inclusion of small plagioclases in the titanaugite; the twinning of the latter; and euhedral basal sections of nepheline (centre, slight left).

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Scottish teschenites contain some alkali-feldspar in addition to labradorite. Usually it does not amount to more than 10 per cent of the whole rock, but it is a significant constituent, and Johanssen considers that a distinctive name, **glenmuirite**, from the locality in central Ayrshire where the exposures of "teschenite" in the Lugar Sill occur, is justified.

Crinanite, named after a Scottish locality by J. S. Flett (1911) is also a type of analcite-gabbro, closely related to teschenite: it is so close, in fact, that there is little doubt that both names have been misapplied—crinanites have been recorded as teschenites and vice versa. Both names are widely used in official Survey publications. The original crinanites were collected from (presumably) Tertiary dykes in the neighbourhood of Loch Crinan, Argyllshire. Stress was laid, in the description, on their relative fineness of grain and perfect ophitic textures. Mineralogically there is little to choose between teschenite and crinanite: both consist of the same mineral assemblage, although the proportions are somewhat different: in crinanite the very small amount (2 per cent) of zeolites is interstitial to the feldspars; but in teschenite zeolites (chiefly analcite) are more abundant and the plagioclases themselves are analcitized. These differences are not sufficient, in our opinion, to justify two names for the same mineral assemblage. Teschenite has the merit of priority.¹

An allied type in which nepheline displaces analcite is **theralite**, consisting of plagioclase within the usual gabbroic range, clinopyroxene and nepheline, together with sundry accessories. Rosenbusch proposed the name (1887) for the plutonic representative of nepheline-tephrite—a fine-grained volcanic rock consisting of this mineral assemblage. The rock originally described under this name by J. E. Wolff from the Crazy Mountains, Montana, does *not* correspond with nepheline-tephrite as it is rich in orthoclase, and must be classed as malignite.

Typical theralites are not common rocks, but specimens closely agreeing with Rosenbusch's conception of what a theralite should be have been described from the Cordilleras of Costa Rica, the Bohemian alkali province and from Scotland. Perhaps the best known theralite, which figures in many teaching collections, comes from Duppau in Sudetenland. Equally typical are certain late-Carboniferous examples occurring in Ayrshire, notably in the Lugar composite sill, in which theralite is accompanied by teschenite. In Scottish theralites titanite is the most abundant component (35 per cent), while labradorite and nepheline occur in approximately equal amounts (16 per cent each in the Lugar theralite). An amphibole, barkevikite in the Scottish rocks, biotite and olivine

¹ Cf. Walker, F., "The term 'crinanite,'" *Geol. Mag.*, 71 (1934), p. 122.

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ANALYSES OF SYENOGABBROS AND MELASYENITES

	I. Shonkinite	II. Borolanite	III. Missourite	IV Essexite.	V. Theralite.	VI. Kentallenite.
SiO ₂	45.77	48.19	45.08	44.70	44.56	48.00
Al ₂ O ₃	8.94	18.52	15.68	14.50	11.75	12.52
Fe ₂ O ₃	3.63	4.51	4.52	2.95	1.77	8.74
FeO	7.13	1.68	3.09	8.35	8.68	—
MgO	12.96	1.12	4.68	7.82	13.31	15.26
CaO	11.56	10.29	11.75	11.25	10.99	7.94
Na ₂ O	1.40	3.44	2.38	4.10	2.46	3.11
K ₂ O	4.60	8.05	8.42	1.71	1.56	2.68
H ₂ O	1.13	3.45	0.88	1.71	2.20	1.36
TiO ₂	0.76	1.75	1.36	2.34	2.27	0.22
P ₂ O ₅ and other constituents	2.09	—	2.37	0.82	0.28	—
	99.97	101.00	100.21	100.25	99.83	99.83

- I. Shonkinite, Square Butte, Highwood Mountains, Montana. F. F. Osborne and E. J. Roberts, *Amer. Journ. Sci.*, **22** (1934), p. 344.
 II. Borolanite, Am Meallan, Loch Borolan, Ross. W. Campbell Smith, *Geol. Mag.* (1909), p. 154.
 III. Missourite, Type B, Italy. Washington, *Amer. Journ. Sci.*, **14** (1927), p. 33.
 IV. Average of two Scottish essexites; from G. W. Tyrrell, *Geol. Mag.*, **60** (1923), p. 254.
 V. Olivine-theralite, Lugar Sill, Ayrshire. G. W. Tyrrell, *Trans. Geol. Soc. Glasgow*, **21** (1947), p. 185.
 VI. Kentallenite, Kentallen, near Ballachulish, Argyll. J. J. H. Teall, *Rep. Geol. Surv.* (1897), p. 22.

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may all be present in varying amounts, but are not covered by the definition. If olivine is present in significant amounts, there are many precedents for calling the rock **olivine-theralite**.

Kylite (G. W. Tyrrell, 1912), is an olivine-rich melatheralite which is correspondingly poor in nepheline. The name was taken from the Kyles of Bute in Scotland.

Lugarite (G. W. Tyrrell, 1912)¹ is one of the very few unique British rock-types. It takes its name from the stratified differentiated sill at Lugar, Ayrshire, where it occurs as veins cutting the other members of the complex—teschenite and theralite. Quantitatively it is of minor importance; but from the mineralogical point of view it is of outstanding interest. Long prisms of barkevikite and titan-augite are conspicuous, embedded in a greenish grey matrix, originally thought to be altered analcite, together with relics of much-altered labradorite and traces of nepheline. Subsequently fresher material became available for study, through boring operations, and the chief felsic component was proved to be nepheline. The assemblage: nepheline, clinopyroxene, labradorite is theralitic; but this rock is so distinctive, chiefly on account of the barkevikite it contains, that retention of the term lugarite seems justified: it is, in effect, barkevikite-theralite. In lugarite two accessories are conspicuous: abundant prisms of apatite and ilmenite showing stages of conversion into brownish translucent leucoxene (Fig. 44).

Rocks which technically fall in the corresponding medium grain-size group and which have actually been recorded are few in number. Relatively finer-grained parts of a kentallenite intrusion would be distinguished conveniently as **microkentallenite**, and similarly in the case of the other types described above. Tyrrell applied the name teschenite-basalt to the fine-grained marginal facies of the teschenite forming part of the Lugar Sill; but this term seems to imply a rock which is between basalt and teschenite in composition, and a better term, in our opinion, is **microteschenite**.

II. TRACHYBASALTS

Trachybasalts are the fine-grained equivalents of syenogabbros. They are essentially rocks of basaltic aspect combining the distinctive compositional features of both trachytes and basalts. This should be a quite unequivocal definition, but the reader is warned that the term trachybasalt is often applied to rocks other than those conforming strictly with this definition. In particular

¹ Tyrrell, G. W., "A boring through the Lugar Sill," *Trans. Geol. Soc. Glasgow*, 21 (1948), pp. 157-202 with references.

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mugearites¹ are repeatedly recorded as trachybasalts; and on the other hand certain rocks recorded as trachyandesites (banakites) should be classed as trachybasalts as the plagioclase they contain is mainly labradorite, but may be as calcic as anorthite.

Many trachybasalts are undersaturated in the sense of containing olivine, but they may be undersaturated also in respect of their felsic constituents. In a manner of speaking the basaltic feldspar—plagioclase—survives; but the trachytic feldspar—sanidine or anorthoclase—may be eliminated in favour of feldspathoid or allied minerals including nepheline or analcite and leucite.

As regards distribution trachybasalts in variety occur in "continental" suites, for example in the mountains of North America, notably in the Yellowstone Park area, where several varieties were described by Iddings and given Indian names (best forgotten). Washington added to the list of named types in his studies of certain Italian volcanic districts. Trachybasalts are important in Jan Mayen among the later Tertiary lavas; and they also occur and compare closely with Yellowstone Park types in certain of the central African volcanic regions. In Britain banakites occur among the Carboniferous lavas in the Midland Valley of Scotland, in the Kilsyth area.²

The most interesting (and the most elusive) thing about trachybasalts is the significance of their mineral composition. The name trachybasalt and an appreciation of its systematic position in the scheme of classification naturally leads one to think in terms of admixture of a trachytic residuum with basaltic magma. It is generally believed that, as a consequence of fractional crystallization, a trachytic residuum may be produced from olivine-basaltic magma. Such a residuum might be incorporated in a later influx of basaltic magma, forming a hybrid magma from which trachybasalt might crystallize directly. Alternatively in slightly different circumstances the trachyte might be represented by aggregates or clots of alkali feldspar embedded in a basaltic matrix, and obviously of mixed origin. Rocks of this type occur among the trachybasalts of Jan Mayen. If this is a true indication of what

¹ It should be understood, that there is no implication that the naming is *wrong*: it was correct according to the definitions in vogue at the time. Most of the confusion has arisen because of uncertainty of the precise meanings of "andesite" and "basalt." At one time it was customary to define basalts as basic lavas, "basic" meaning a rock containing between 45 and 55 per cent of silica, regardless of its mineral composition. This permitted lavas containing plagioclase of *any* composition to qualify as basalts: hence, following H. S. Washington, there were albite-basalts, oligoclase-basalts, andesine-basalts, labradorite-basalts and bytownite-basalts. Mugearites are basic, they contain oligoclase and were hence classed as oligoclase-basalts; or if the writer was impressed by their characteristic "trachytic" (fluxional) texture, they were recorded as trachybasalts.

² "Geology of East Lothian," *Mem. Geol. Surv.* (1910), p. 131.

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happens, these rocks should display a wide variation in the proportion of basaltic to trachytic material in different types. One of Washington's types—vulsinite—contains over 70 per cent of Na-orthoclase (the trachytic contribution) as against anorthite phenocrysts 6 per cent, and 12 per cent of labradorite in the groundmass; so that this type is three-quarters trachyte and only one-quarter basalt.

Undersaturated trachybasalts include a number of rock-types bearing a superficial resemblance to basalts: they are dark-coloured, fine-grained rocks occurring either as lava flows or in dykes. As in certain basalts, phenocrysts of augite and olivine may be visible even in hand specimens; but there the resemblance to basalts ends, and the differences are much more significant than the resemblances. The feldspathoid minerals nepheline (or the chemically similar analcite) and leucite occur, in place of the alkali-feldspar of the saturated trachybasalts.

Classification is based upon the kind of feldspathoid mineral present, and the presence or absence of olivine.

Olivine-free types are termed **tephrites** (Cordier, 1816) of which there are three varieties, distinguished by prefixing nepheline-, leucite- or analcite- as circumstances demand. The addition of olivine by analogy with many other rock-types should give olivine-tephrite; but long-established precedent demands that they shall be called **basanites** (Brongniart, 1813), again with the appropriate mineral prefix giving nepheline-, leucite- and analcite-basanite. It would be not inappropriate if varieties containing only *small* (accessory) amounts of feldspathoids were termed nepheline-basalt, leucite-basalt or analcite-basalt respectively; but unfortunately these names have been wrongly applied in the past to rocks which are not basalts as they contain no plagioclase.

With regard to the details of mineral composition and texture little need be added to what has already been written. Thin sections convey the impression that one is looking at basalt, with nepheline and/or leucite added. The rocks are generally strongly porphyritic. Probably the most widely distributed variety is leucite-basanite, typically represented by well-known lavas from Monte Somma and Vesuvius, containing phenocrystic leucite, olivine, augite and plagioclase, embedded in groundmass which may include all of these minerals as second generation microlites, with magnetite the commonest accessory. In the nephelinic varieties nepheline may occur as easily recognized phenocrysts, but when it occurs only in the groundmass it is less readily identified. Of the analcite-bearing varieties only analcite-basanite is at all common, and is, in effect, olivine-basalt with a variable amount of analcite interstitial to the groundmass microlites.

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As regards origin, all that was said above about trachybasalts in general applies also to these rocks. The trachytes, which are regarded as ultimate differentiates from olivine-basalt magma, are often somewhat undersaturated and range from orthotrachytes (saturated) through nepheline-trachytes to phonolites. By analogy the trachytic portion of a trachybasalt may be represented by alkali-feldspar (sanidine or anorthoclase) or its place may be taken by nepheline or leucite, according to the balance between K' and Na' .

The close mineralogical analogy between tephrites and especially basanites on the one hand and teschenites, theralites and essexites on the other will be apparent to the reader; and again it will be noted that there are no precise analogues of leucitic varieties among the rocks in the coarse-grain category because leucite cannot survive the deep-seated conditions of crystallization.

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RELATED TYPES

UNDER this heading are included several rock-types which superficially resemble basalts: they are dark in colour, fine in grain, basic to ultrabasic in composition and occur either as lava-flows or as petrographically identical dyke rocks. Phenocrysts, often visible in hand specimens, include olivine and augite, heightening the resemblance to basalts; but in other types megascopic leucite may be a distinctive feature, while in thin sections other feldspathoids are visible.

Classification is effected on the basis of the kinds of feldspathoids present, the presence or absence of plagioclase, and of olivine. In the first instance distinction is made between plagioclase-bearing and feldspar-free varieties. This distinction is fundamental and cannot be over-emphasized. The plagioclase-bearing varieties—tephrites and basanites—were discussed above as under-saturated trachybasalts. Examination of thin sections of these rocks conveys a strong impression that they are fundamentally basaltic, with a variable amount of feldspathoid minerals added: tephrites and basanites *without* their nepheline, analcite or leucite are just basalts or olivine-basalts respectively.

No such impression is conveyed by the feldspar-free types: compositionally they have nothing in common with basalts as is clearly indicated in the varieties illustrated. The absence of plagioclase automatically excludes these lavas from the basalt suite (the right-hand column of the table on p. 247) and also from the middle column—the trachyandesite-trachybasalt suite. With more silica available in the magma leucite, kalsilite and nepheline would have been represented by alkali-feldspar, therefore the affinities of these rocks are with the syenite-microsyenite-trachyte series. They are compositionally allied to phonolites and leucitophyres, with which they are sometimes closely associated in the field.

It may be useful to remember that we are dealing with quenched rocks containing unstable mineral associations. This is particularly true of the kalsilite- and leucite-bearing varieties, the coarse-grained equivalents of which would be found among the shonkinites. The corresponding nephelinic lavas are analogous with the ijolite-melteigite series.

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Difficulties of nomenclature arise when the feldspar-free lavas are concerned. Among the rock-names in common use some will probably shock the reader as being illogical and conveying entirely false impressions of the mineral composition. This is particularly true of the terms "nephelinite," "leucitite" and "melilitite." If it were possible to start again there would be no difficulty in finding suitable names; but a hundred and fifty years ago when these names were first introduced, petrographic nomenclature was in its infancy; the urtite-ijolite-melteigite series was unknown, so too were the various monomineralic rocks upon the logical naming of which a good deal depends. We have already established the principle that the rocks which consist of one essential mineral only should take the name of that mineral with the addition of the uniform termination "-ite." Thus albitite, pyroxenite, bronzitite, hornblendite, etc., are established terms of this kind. But when the names "nephelinite" and "leucitite" were applied to varieties of plagioclase-free feldspathoidal lavas, it was not anticipated that they would one day be required for the rock consisting solely of nepheline or leucite as the case might be. Now, however, the (nearly) pure nepheline rock is known, as the end-member of the urtite-ijolite-melteigite series, and there is no alternative to the name nephelinite. The facts appear to speak strongly for themselves. It is inconsistent with the rules to which we are working to continue using these names in an inadmissible sense and we do not propose to do so, but to attempt to introduce a new set of terms which roll easily off the tongue, and which adequately suggest the mineral composition of the rocks concerned. They will prove acceptable, it is believed, to the general geological reader. These things can be said of very few rock-names. To avoid confusion we propose to bracket the old with the new terms.

All the rocks under consideration are mafic, and the feldspathoid minerals are diagnostic of the different varieties. Both facts may be combined in the rock names as shown in the table below:

Mineral composition.	Suggested names	Names in current use.
Kalsilite plus mafic minerals	Kalmafitite	Mafurite, katungite
Leucite plus mafic minerals	Leumafite	Leucitite, ugandite
Nepheline plus mafic minerals	Nemafite	Nephelinite
Melilite plus mafic minerals	Melmafitite	Melilitite, Melilitite-basalt

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It may be helpful to tabulate the nomenclature and mineral compositions of the various plagioclase-bearing feldspathoidal lavas already described, and to compare these with the plagioclase-free types. Varietal names are also suggested for the latter to cover combinations of two or more feldspathoids and the possible addition of olivine.

Group 1: Plagioclase-bearing feldspathoidal lavas

The fundamental definition: plagioclase (An₅₀) plus pyroxene and accessories = BASALT.

The basaltic assemblage plus essential feldspathoid = TEPHRITE.
Varieties named according to the dominant feldspathoid:

Nepheline-tephrite
Analcite-tephrite
Leucite-tephrite.

The tephritic assemblages plus additional olivine = BASANITE.
Varieties as for tephrites:

Nepheline-basanite
Analcite-basanite
Leucite-basanite.

Group 2: Plagioclase-free, therefore NOT basaltic lavas,

Kalsilite plus mafics, with accessories = KALMAFITE.

Kalsilite dominant with (say) nepheline:

Nepheline-kalmafite

With additional olivine:

Olivine-nepheline-kalmafite.

The corresponding type containing dominant nepheline = NEMAFITE

Olivine-nemafite
Leucite-nemafite
Olivine-leucite-nemafite.

With dominant leucite the corresponding term = LEUMAFITE

Olivine-leumafite
Nepheline-leumafite
Olivine-nepheline-leumafite.

Of the many rock-types involved the leucitic varieties are the most familiar and the most easily identified on account of the

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distinctiveness of leucite in thin section as well as in some of the hand specimens. In the following account we attempt to cover the necessary petrographic detail and the facts of distribution together instead of separating the two aspects.

Mafic leucitic lavas

The most familiar leucitic lavas occur in the Roman Volcanic Province which includes Vesuvius and Monte Somma. At one time they were thought to be restricted to this region; but they are now known to occur in all the continents, notably in the aptly named Leucite Hills, Wyoming, U.S.A., the Kimberley District in Western Australia, the Dutch East Indies, Antarctica and several parts of Africa including the neighbourhood of the Western Rift, the former Belgian Congo, Uganda and adjacent parts of Tanganyika.

The central and eastern African region is by far the most important of these. The different occurrences are being actively explored and studied at the present time and knowledge concerning them is growing fast. The leucitic lavas include leucite-tephrites, leucite-basanites, the less basic leucitophyres and phonolites, as well as plagioclase-free leumafites ("leucitites"). Some of the latter are associated with and grade into kalsilite-bearing lavas; while nepheline occurs, too, in an accessory capacity. An augite-rich olivine-leumafite (termed "ugandite" by Holmes, 1945) is an uncommon variety which grades into leucite-melabasanite with the incoming of plagioclase.

With regard to the Roman Volcanic Province it may truthfully be claimed that the first feldspathoidal lavas to be recognized as such came from the neighbourhood of Vesuvius, and they remain the main source of supply of teaching material. The most distinctive types are typical **leucite-basanites** in which all the essential constituents occur as phenocrysts. The off-white icositetrahedral leucites occurring in these rocks are particularly distinctive and well known. Large first-generation leucites display their characteristic polysynthetic twinning in thin sections, but the smaller crystals of the second generation are often singly refracting. The small leucites illustrated in Fig. 118 are of more than usual interest on account of the minute glass inclusions which they contain. In their arrangement they are identical with those described by Bowen and Morey from leucites obtained by quenching during their classical experiments on the melting of orthoclase. As will be seen from the illustration, the inclusions are often significantly grouped in threes or fours about the points marking the emergence of triad or tetrad axes of symmetry.

A unique leucitic rock from the Roman Volcanic Province was

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named *italite* by Washington (1920),¹ though leucite would have been accurate and informative, as leucite is the only identifiable mineral in the rock. It occurs as blocks in an agglomerate in the Alban Hills and consists of large euhedral leucites cemented together by a small amount of whitish material which may be altered



MKW

FIG. 118

Leucumafite ("leucite") Capo da Bove, near Rome.

Small euhedral leucites, clinopyroxene, a little mica, accessory apatite and magnetite. The groundmass is colourless glass.

glass. If it is, the rock must be regarded as a phenocryst-rich extrusive. Evidently *italite* is an accumulative rock, and there can be little doubt that it represents a roof- (as distinct from a floor-) accumulation of sorted phenocrysts. As we have argued in the case of other monomineralic rocks, anorthosite for example, such an accumulation could neither be intruded nor extruded as lava, but would almost inevitably be disrupted by volcanic vent explosions, as happened in this instance.

¹ Washington, H. S., "Italite, a new Leucite Rock," *Amer. Journ. Sci.*, 1 (1920), p. 33.

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Leumafites in variety occur in the West Kimberley area in Western Australia.¹ Four types connected by intermediate varieties have been described, emphasis being placed on the mafic minerals which they contain. Phlogopite is perhaps the most significant as being equivalent to leucite and olivine combined. This Mg-rich mica makes up to a quarter of the whole rock in one type; it is accompanied by a manganese-amphibole of rare type in another, while a third contains prisms of clinopyroxene and olivine pseudomorphs. It is therefore an olivine-leumafite ("olivine-leucitite"). The most interesting fact about these Australian rocks is that, although they are described as the most richly leucitic rocks in the world, chemically speaking they should contain no leucite as there is plenty of silica available in the rocks to have converted it all into sanidine. The silica needed for this conversion is present in a copious glassy base, but was made inoperative by quenching. One chemical feature of the magma from which these lavas crystallized is a high content of titanium: much of it is present in the mica, but some occurs as the uncommon accessory, rutile, which amounts to as much as 6 per cent in one type.

Several varieties of leucitic lavas occur in the Leucite Hills, Wyoming.² Most of them are leucitophyres; but one of them, on account of its mafic character, may be considered here, especially in view of what has been stated in the immediately preceding paragraph. The rock was termed madupite by Cross (1898). It consists of diopsidic pyroxene making up almost half the rock, phlogopite comes second in importance, while the distinctive accessory, perovskite, also occurs. It will be noted that leucite does not occur in the list of minerals; but up to one-third of the rock consists of glass having the composition of leucite, so that *potentially* leucite is present. The rock is named and classified on the basis of a mineral composition that it would have had, if crystallization had not been arrested by quenching. Whether or not this is a sound principle is a debatable point. The rock would be described as a phlogopitic leumafite ("leucitite") containing leucite-glass. A short phrase is needed to convey the right impression of the nature of this rock: the single word "madupite" certainly does not.

Mafic kalsilite-bearing lavas

These are the rarest of lavas, at present known to occur in only two areas in Africa. The mineralogical characteristics of kalsilite

¹ Wade, A., and Prider, R. T., "The Leucite-bearing Rocks of the West Kimberley Area, Western Australia," *Quar. Journ. Geol. Soc.*, **96** (1940), p. 39.

² Cross, W., "The Igneous Rocks of the Leucite Hills, etc.," *Amer. Journ. Sci.*, **4** (1898), pp. 126, 134 and 139.

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and its relationship to other feldspathoids have been already discussed. Although these rocks are feldspar-free and ultrabasic in composition they are particularly noteworthy on account of the variety of unsaturated minerals which they contain—felsic, mafic and accessory. The several varieties have been named after the type-localities.

An olivine-kalsilite termed mafurite by Holmes (1945)¹ contains small phenocrysts of olivine and augite set in a matrix of green augite and accessories including abundant perovskite, with interstitial kalsilite. Had there been a little more silica available in the magma kalsilite would have given place to leucite and the mineral assemblage would have been olivine, augite, leucite and accessories, already encountered as olivine leumafite ("olivine-leucite").

A very distinctive variety named katungite after the extinct volcano Katunga, in Uganda, contains kalsilite, melilite, olivine, perovskite, titanomagnetite and interstitial glass. With slight increase in silica katungite grades into varieties less completely undersaturated and containing leucite instead of kalsilite and augite instead of melilite. This variety, distinguished as ugandite by Holmes, is particularly rich in mafic minerals, otherwise the mineral assemblage is that of olivine-leumafite (*i.e.*, it is a melano-cratic "olivine-leucite").

Kalsilite-bearing lavas of exceptional interest have been discovered in the Belgian Congo.² In general they are close to katungite—the same minerals occur in both, but with nepheline most intimately associated with kalsilite in the Congo varieties. These two minerals form glomerophytic aggregates of phenocrysts which show varying degrees of unmixing of the two components. The distribution of alkalis in these lavas makes an interesting study. Thus there is a reciprocal relationship between the amount of KAlSiO_4 in the nephelines, and the quantity of leucite in the rock: in specimens relatively poor in leucite, the associated nephelines are rich in K' -ions; but if the rock is leucite-rich the accompanying nephelines are deficient in K' -ions.

In the permanent magma lake of Nyiragongo in the Congo the accumulation of aggregates of phenocrysts swept up by turbulent currents in the magma may be observed. This is a feature of special significance in connection with the formation of "accumulative" rocks including phenocryst-rich lavas which display a condition of unstable equilibrium between phenocrysts and magma now represented by groundmass, as described under "Andesites."

¹ Holmes, A., "A suite of volcanic rocks from S.W. Uganda," *Min. Mag.*, 26 (1942), p. 197; also Holmes, A., and Combe, A. D., "The kalsilite-bearing lavas of . . . S.W. Uganda," *Trans. Roy. Soc. Edin.*, 61 (1945), pp. 359-79.

² Sahama, Th. G., "Kalsilite in the lavas of Mt. Nyiragongo, Belgian Congo," *Journ. Petr.*, 1 (1960), p. 146.

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Mafic nepheline-bearing lavas

Nepheline-bearing varieties comprise the commoner nepheline-tephrites and basanites and the much rarer nematicites ("nephelinites") and olivine-nematicites ("olivine-nephelinites"). The nematicites at their best are very distinctive rocks. Nepheline may occur in two generations, both euhedral, or the second generation, instead of forming the characteristic small square and hexagonal sections as in Fig. 119, may be interstitial to the other constituents and is then less easy to identify. Indeed a staining test or even X-ray analysis may be necessary before the mineral is positively identified. The nematicite drawn in Fig. 119 is a striking example of its kind, with the nephelines abundant and euhedral. Mafic minerals include large titanagites with pale green outer zones and extensive poikilitic phlogopitic mica. Of the accessories sphene is abundant, and its relationship to the other minerals is unusual: some of the crystals are euhedral, but in the main it forms irregular crystal grains poikilitically enclosing the groundmass microlites, especially nephelines. This sphene is evidently late in the crystallization sequence. Two textural features call for comment: the zoning of the pyroxenes and the late growth of the outer zones so that nephelines are enclosed in them. This is true also of the phlogopite crystals.

All the nephelinitic lavas referred to above contain identifiable nepheline; but there are in addition certain rocks of the same chemical composition which contain "occult" nepheline: *i.e.*, potential nepheline represented by glass of the same composition. If the nepheline had been *actual* instead of *potential* the rocks would have been classified as nepheline-basanites; but it is customary to distinguish them as **basanitoids**. They are, in effect, nepheline-glass basanites.

Melilite-bearing lavas and dyke rocks

Melilite, like the feldspathoid minerals, may "proxy" for feldspar in undersaturated rocks, particularly those which have been quenched. It differs fundamentally of course, as it contains Ca" (and often Mg") instead of K' and Na' and therefore takes the place of plagioclase instead of alkali-feldspar. Until the discovery of the alkali-rich lavas in parts of Africa, notably South-West Uganda and the Belgian Congo, melilite-bearing lavas were represented by one type only which was miscalled "melilite-basalt", and later "melilitite."

The original **melmafite** ("melilite-basalt," Hochbohl, 1883) occurs in the Swabian Alps. It is thoroughly ultrabasic, with a silica percentage of only 34. Nearly half the rock is olivine, approximately 40 per cent is melilite, while small quantities of nepheline,



FIG. 119

Nemaite (nephelinite), summit of Lodwar Hill, northern Turkana, Kenya Colony.

For description see text, p. 364. Pyroxene, stippled; magnetite, black; sphene, heavy stipple. Phlogopite with cleavage indicated. There are two generations of pyroxene.

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perovskite, apatite and calcite also occur. "Calcitic nepheline-melmafite" concisely describes the type rock. Similar lavas, with or without olivine, occur as rarities in several localities including Tasmania, but more commonly in Africa including Namaqualand and the central African volcanic region in Kenya and South-West Uganda. In the latter area melilite is accompanied by kalsilite in the variety termed **katungite** by A. Holmes, already described under "kalsilite-bearing types."

Although the rock called **alnöite**, from Alnö, an island off the coast of Sweden, was originally classed as melilite-basalt, the name was changed by Rosenbusch to alnöite, *on the grounds that it is intrusive*. Although we strongly disagree with Rosenbusch's reason for changing the name, alnöite is sufficiently distinctive in itself to justify its name. Fundamentally the type is one-third each of melilite and biotite, the remainder of the rock being made up of pyroxene, calcite and olivine in order of importance, with various minor accessories. At Isle Cadieux¹ between Montreal and Ottawa, a small intrusive complex of alnöitic rocks occurs. Two types in particular are outstanding: one contains the mineral monticellite in addition to ordinary olivine (chrysolite); while the other, occurring as mere streaks in the alnöite, consists almost wholly of biotite and melilite.

Alnöite has been recorded also from certain of the African ring-complexes associated with carbonatites, notably at Chilwa, Nyasaland. Actually in this rock melilite is represented by carbonate pseudomorphs.

As a matter of convenience certain melilitic rocks of coarse grain may be included here. Three varieties have been described under the names **turjaite**, **okaite** and **uncompahgrite**.

Turjaite (Ramsay, 1921) forms a small intrusion at Turja in the Kola Peninsula. In hand-specimens large crystals of titaniferous biotite are conspicuous, embedded in a matrix of melilite which makes up nearly half the rock. Nepheline, too, is an essential component, while apatite (very abundant in some facies of the rock), magnetite, and, as might be expected, perovskite, are important accessories. Although olivine and melanite garnet are usually rare, they also become important in some specimens. The frequent occurrence of calcite, as in certain nepheline-syenites, and the association of melilite-bearing rocks with carbonatites are significant facts bearing on the genesis of this rock.

Of still more extraordinary composition is the rock, also judged from its grain-size to be deep-seated, termed **Okaite** by Stansfield. In its mineral composition okaite is very close to turjaite, but melilite is slightly more abundant, biotite less so, while the place of

¹ Bowen, N. L., "Genetic Features of Alnöitic Rocks at Isle Cadieux, Quebec," *Amer. Journ. Sci.*, 3 (1922), p. 1.

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nepheline is taken by hauyne. The type-rock was discovered in the Oka Hills, near Quebec, and, although a great rarity, it is noteworthy as having a particularly low silica percentage, only 29, while it contains 25 per cent of CaO.

A very rare coarse-grained melilitic rock has been recorded under the name uncompahgrite from a locality in central Africa. The most conspicuous feature is titaniferous biotite occurring as large reddish brown crystals, together with aggregates of titanomagnetite, perovskite and apatite, all embedded poikilitically in melilite. Thus essentially the mineral composition is biotite plus melilite—the same as that of the rock occurring in the Isle Cadieux complex, noted above. It also is a melilite-biotitite, carrying accessories which seem to be characteristic of melilitic rocks.

Finally reference may be made to a British occurrence described by C. E. Tilley, from Scawt Hill near Belfast, where a plug of Tertiary dolerite penetrates chalk. Both nepheline and melilite occur in a hybrid zone formed by interaction between the magma and the carbonate rock.

With regard to the origin of the melilitic rocks the following facts appear to be significant:

(1) the close association of many, but not all, the rocks concerned with carbonatites (Alnö and Chilwa) and the occurrence of accessory primary calcite in some of them;

(2) the exceptionally low silica percentage and the extremely high lime-content, which is the highest in all igneous rocks excluding carbonatites;

(3) experimental studies have proved that at high temperatures nepheline and augite combine to form olivine and melilite with monticellite as an intermediate product of the reaction;

(4) when calcium carbonate is added to fused basalt, melilite crystallizes from the melt; and

(5) melilite occurs in the hybrid reaction zone between chalk (CaCO_3 -rock) and basalt at Scawt Hill, Co. Antrim.

Of these facts we would emphasize (3), (4) and (5) as being specially significant. The association of nepheline and augite is one of the commonest in the nephelinic feldspathoidal lavas. The experimental results suggest that this association is essentially the same as melilite with olivine, encountered far less frequently. This indicates that physical conditions, rather than exceptional composition, determine whether the association nepheline-augite, or melilite-olivine will separate from the magma in any given case. The latter association is characteristic of quenched rocks, while the former is the more stable combination, and that which is much more frequently

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encountered. However, chemical factors must also play a part. The evidence suggests that desilication of basic magma by reaction with carbonates—whether as limestones or carbonatites is immaterial—probably plays a critical part. Thus the origin of melilite may be closely analogous to that of nepheline, as discussed in the following chapter.

Finally reference must be made to those mafic lavas which contain analcite. The only variety which calls for recognition is **analcite-basanite**, comparable with nepheline-basanite, but with analcite substituted for nepheline. Thus the mineral assemblage comprises olivine, pyroxene, basic plagioclase, analcite and accessories. This is the mineral assemblage of the Scottish essexites and teschenites which are the coarse-grained equivalents of these analcite-bearing lavas. It will be noted that it is also the mineral assemblage of normal olivine-basalt, with analcite in addition; and the reader is warned that small quantities of analcite are easily overlooked in a thin section, especially when it is interstitial. A British example occurs at Calton Hill, Derbyshire.¹ The rock is very similar in general appearance to olivine-basalt, but the analcite is rather more obvious than usual as it forms small spherical ocelli.

¹ Tomkeieff, S. I., "The volcanic complex of Calton Hill, Derbyshire," *Quar. Journ. Geol. Soc.*, **84** (1928), p. 715.

ORIGIN OF THE FELDSPATHOIDAL ROCKS AND CARBONATITES

HAVING in the preceding chapter concluded our review of the petrography of the many varieties of feldspathoidal rocks, it remains to consider the evidence relating to the origin of some of them. The diversity of rock-types is so great that obviously no single mode of origin can cover them all, and the discussion which follows is necessarily selective. It is mainly the strongly alkaline rocks rich in feldspathoids with which we are concerned and which provide the major problems. Three aspects of the problems of origin are examined here:

- (1) the origin of nepheline-syenites and related feldspar-free types (the urtite, ijolite, melteigite series);
- (2) the nature and origin of the closely associated carbonatites;
- (3) the relationship between the coarse-grained feldspathoidal rocks and lavas of equivalent composition.

Items (1) and (2) are discussed from the point of view of visible field relations; but the origin of the magmas represented by the feldspathoidal lavas is a much more elusive problem which is, of course, highly speculative and is therefore only touched upon. For reasons that will emerge we deal separately with nephelinitic and leucitic lavas.

(1) *The origin of nepheline-syenites and related feldspar-free types*
The majority of nepheline-syenites are closely associated with, and apparently intrusive into, granitic rocks. At one time this association was interpreted as demonstrating that the former had been produced from granitic magma by a process of desilication. The desilicating agent was thought to be limestone which in many occurrences was closely involved with the granite and nepheline-bearing rocks. R. A. Daly formulated a hypothesis according to which reaction between limestone and granitic magma would produce Ca- and CaMg-rich silicates which, on account of their high specific gravity, would tend to sink, leaving a desilicated magma-fraction from which nepheline could crystallize. Although Daly proposed the

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idea and gave it wide publicity, S. J. Shand had a major share in developing the hypothesis largely as a result of his detailed studies of alkali complexes including that at Loch Borolan in Assynt and several in Africa, especially Spitzkop and Palabora. These and other examples are discussed in considerable detail in his *Eruptive Rocks* which all students interested in feldspathoidal rocks should study.

In the following discussion three alkali complexes are considered: all three involve "granites" and carbonate rocks and have been used to illustrate the Daly-Shand hypothesis. The first two are situated in the Baltic Shield: the Alnö Complex¹ forms part of an island off the Swedish coast, while the Fen Complex² lies just west of the Oslo Graben in southern Norway. Both are ring-complexes, the alkali rocks enclosing central masses of carbonate rocks originally thought to be sedimentary and termed limestone, but now known to be carbonatite. Very poor exposure is a characteristic of many alkali-complexes, including these two, and must be held responsible for the early misinterpretation of the field relationships. The two complexes are so similar in their essential features that they may be described together. Both are set in "granitic" environments, the country-rock being granite-gneiss forming part of the Basement Complex. The central carbonatite forms a plug about which the other rock-bodies are arranged concentrically. They comprise (1) an inner zone of feldspathoidal rocks in variety; (2) a zone of nepheline-syenites formed by metasomatism of the gneiss *in situ*; (3) an outer zone of saturated syenites also resulting from metasomatic alteration of the gneiss; and (4) a zone of fractured gneiss.

The first stage in the formation of the complex involved the explosive fragmentation of the rocks above an advancing plug of magma. Progressive metasomatism of the brecciated rocks produced the outer zone of syenites by elimination of free quartz, or rather its incorporation in newly crystalline silicates which include a distinctively coloured amphibole, though aegirine is the characteristic mafic mineral. No feldspar was actually introduced into the gneiss, though new alkali-feldspars of both sodic and potassic types were developed. To what extent they owe their origin to redistribution of material already in the gneiss it is impossible to say because the original gneiss is not available for analysis. Some ions must have been introduced. The new potassic feldspar contains Ba'', so that in a manner of speaking it contains celsian.

¹ Von Eckermann, H., "The alkaline district of Alnö Island," *Sver. geol. undersök.*, No. 36 (1948), p. 176.

² Saether, E., "The alkaline rock province of the Fen Area in Southern Norway," *Kong. Norske Vidensk. Skr.*, No. 1 (1957). See also summary account in "Geology of Norway" (Oslo, 1960), edited by Holtedahl, Olaf.

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Desilication beyond the point where free quartz was eliminated gave rise to nepheline-syenites which were formed from solid gneiss by metasomatism in place. This is proved by the fact that

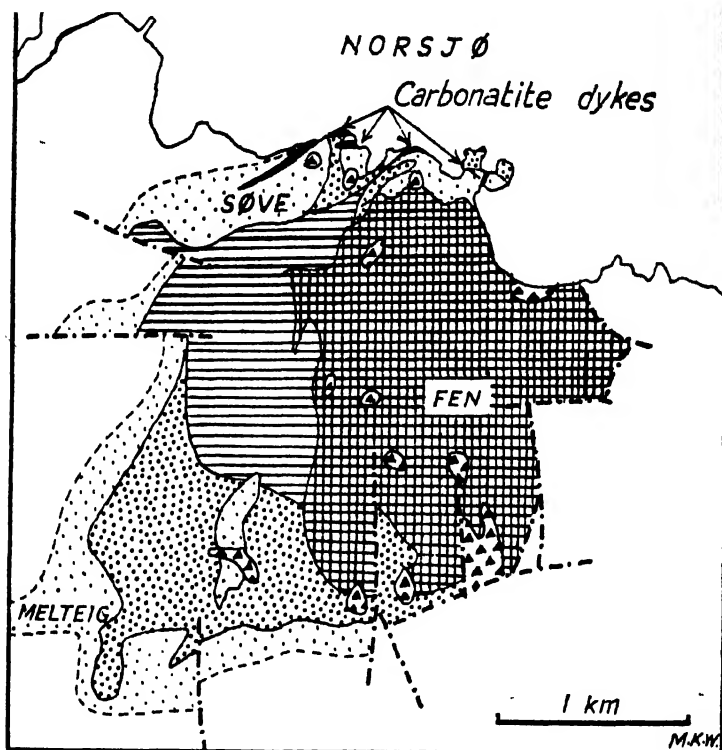


FIG. 120

The Fen alkaline igneous complex (after E. Saether).

Pre-Cambrian left blank; fenites (metasomatic syenites and nepheline-syenites), light stipple; rocks of the ijolite-melteigite series, heavy stipple; mixed carbonate-silicate rocks, horizontal ruling; calcite, dolomite, ankerite and haematite rocks (in part metasomatic), cross-ruling; carbonatite dykes, black; kimberlite breccia-vents, triangular ornament.

structural features may be traced from the gneiss into the nepheline-syenites. The latter, therefore, are nepheline-syenites in composition though they are definitely not magmatic in the strict sense. Brøgger¹

¹ Brøgger, C. W., "Die Eruptivgesteine des Kristianiagebietes—IV. Das Fengebiet in Telemark, Norwegen," *Vidensk. Skr. 1, Math. Nat. Kl.*, No. 9 (1920), p. 1921.

This is a classic monograph containing the original descriptions of a large number of feldspathoidal and carbonate rocks to which, unfortunately, he gave an equally large number of names of the type-locality variety, among them melteigite, fenite and sövite.

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introduced the term *fenite* for rocks of this nature, while the term *fenitization* is widely used for the metasomatizing processes involved: the latter, as we have seen, resulted in the progressive elimination of silica (desilication) and the accession of alkalis. Lying inside the encircling fenites occur feldspathoidal rocks including nepheline-syenites and feldspar-free types belonging to the urtite-ijolite-melteigite series. These rocks form a major part of both the Alnö and Fen Complexes. No relict structures inherited from pre-existing rocks are visible, and therefore there is no evidence suggesting derivation from the latter. On the contrary they exhibit sharp and cross-cutting contacts with adjacent rocks and are sometimes clearly intrusive. Further, some of these rock-types occur also in radially disposed dykes which are regarded as magmatic. One view concerning the origin of these apparently magmatic nephelinitic rocks is that, after metasomatism, the material was rendered mobile (mobilization is the term commonly used for the process) and thus achieved a magmatic condition. Thus the metasomatic stage is a prelude to the formation of magmatic rocks.

The sequence of events outlined above provides the general pattern for a large number of similar complexes, especially in Africa. Naturally they differ in detail. The country-rock is Basement Gneiss in many, but not in all cases: it may be sandstone or rhyolite tuff; but in most cases desilication like that observable at Alnö and Fen has led to the formation of syenites and/or nepheline-syenites. Fenites are frequently, though not invariably visible; while in one case of outstanding significance, denudation has bitten into the rocks less deeply than usual with the result that the remains of an extrusive phase including lavas and tuffs have survived, as well as the more usual coarse-grained rocks of the intrusive phase.¹ The importance of these relationships will not be lost on the reader, of course.

In the Haliburton-Bancroft area in Ontario a very different type of occurrence of nepheline-syenites is found. The rocks form an integral part of the metamorphic complex of the Canadian Shield. It is a classic region; but once again the field relations are very differently interpreted. As compared with Alnö and Fen, the fundamental difference is the association of granites with nepheline-syenites and limestones of sedimentary origin—the very extensive Grenville Limestone. Formerly the nepheline-syenites were regarded as having originated by desilication of granitic magma in the manner postulated by Daly and Shand; but Tilley has made a detailed study of some of the nepheline-bearing rocks, and has shown that the granites were intruded after the formation of the former, so that

¹ King, B. C., "The Napak area of Southern Karamoja, Uganda: a study of a dissected late Tertiary volcano," *Geol. Surv. Uganda*, Mem. 5 (1949).

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the limestone-syntaxis hypothesis may not apply.¹ The nephelinic rocks are again interpreted as being metasomatic; but the process involved silication of the limestone which has been converted into varied assemblages of plagioclase and nepheline associated with hornblende, biotite, microcline and other minerals. The process whereby the nephelinic rocks were generated (aptly termed nephelinization²) was clearly metasomatism of the reverse kind to that which was operative at Alnö: there desilication of "acid" rock took place; but at Bancroft silication of very basic material occurred. In both areas nephelinic rocks were the products of metasomatism *in situ*, but nepheline-syenite magma was available, too, though whether it should be regarded as cause or effect is a baffling problem.

It is noteworthy that, of the materials needed to build-up nepheline, all were originally present in the granite-gneiss at Alnö—they only needed redistributing; but at Bancroft none of the necessary cations were present—they all needed to be introduced from outside. The source of these is problematical: Tilley considers it probable that they were derived from a basic magma in depth; but granitic magma was on the spot, so too was another magma represented by the magmatic nepheline-syenites: their interrelationship is still quite unknown.

(2) *Carbonatites*

In the foregoing account passing reference has of necessity been made to the more or less pure carbonate rocks now widely known as carbonatites.³ In all known occurrences they form parts of volcanic complexes dissected to different depths and invariably associated with feldspathoidal rocks. In the most striking instances the carbonatite forms a central plug rising prominently above the lower ground occupied by the associated silicate rocks. The greater resistance of the carbonatite to weathering under the climatic conditions obtaining in central and East Africa is a particularly fortunate circumstance: the characteristic relief of these carbonatite, nepheline-syenite complexes has enabled the Overseas Geological Surveys to locate many from aerial photographs.

¹ Tilley, C. E., "Problems of alkali rock genesis," *Quar. Journ. Geol. Soc.*, **108** (1958), p. 323.

This valuable review consists of two parts: the first deals with general problems; the second is a petrological study of an occurrence of nephelinic rocks east of Bancroft.

² Gummer, W. K., and Burr, S. V., "Nephelinized paragneisses in the Bancroft Area, Ontario," *Journ. Geol.*, **54** (1946), pp. 137-68.

³ Campbell Smith, W., "A review of some problems of African carbonatites," *Quar. Journ. Geol. Soc.*, **112** (1956), pp. 189-220; also King, B. C., and Sutherland, D. S., "Alkaline rocks of eastern and southern Africa," *Sci. Prog.*, **48** (1960), Pt. I, pp. 300-21, and Pt. II, pp. 504-24.

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In addition to the central plugs, carbonatites form dykes radiating outwards from the volcanic centres, and also cone-sheets concentric with the latter. In all three modes of occurrence carbonatites are comparable with, say, dolerites, and a magmatic origin is inevitably suggested by the demonstrable field relations. The carbonatite cone-sheets at Alnö, for example, are exactly comparable with the doleritic cone-sheet complexes in the British Tertiary volcanic province.

In spite of these structural analogies there has been widespread reluctance to accept carbonatites as magmatic for two chief reasons: (1) the high melting point of calcite ($1,339^{\circ}\text{C.}$); and (2) the very high vapour pressure of CO_2 which it is thought must have been involved. With regard to the latter, it is significant that the intrusion of carbonatites has been accompanied by intense brecciation of the country-rock, as for example at Alnö, proving that gas pressures were indeed very high. It may be mentioned in passing that the suggestion has been made, by Bowen among others, that carbonatites might be replacive after silicate rocks; but this is generally not supported by the features displayed in thin sections. Critics of the carbonatite-magma hypothesis emphasize the point that observed intrusive relationships, which they do not query, do not prove a magmatic origin. That is agreed. In Persia in particular rock-salt has punched its way through thousands of feet of sedimentary rocks, developing a strong cross-cutting relationship towards them in so doing. It has been extruded like lava from cones aping those of genuinely volcanic origin; and underground has formed intrusions to which such terms as laccoliths might be applied. In spite of these facts nobody finds it necessary to invoke a rock-salt magma. On account of the ease with which plastic deformation and flow may be induced in calcite, there is no doubt that in a suitable environment calcite could behave like rock-salt. Actually the two cases are not analogous. Where the rock-salt originates and how it reaches the surface are reasonably well understood; but where and how carbonatite originates is still a complete mystery, and its almost invariable association with nepheline-syenites and related rocks is not understood.

Viewed simply as rock-types, carbonatites are very variable. Some are pure carbonate rocks consisting of coarsely crystalline calcite, dolomite or ankerite in different instances. At Alnö, for example, one set of cone-sheets is calcitic while a second set, converging on a different focus, is dolomitic. Apatite and magnetite are frequent accessories, and there is a gradation into silicate rocks. At the one extreme are such rocks as that which has been given a distinctive type-name (tuvinite, Yashina, 1957) and consists of nepheline 75 to 95 per cent and calcite 5 to 25 per cent. Nepheline-syenites of several

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types contain small amounts of accessory calcite, undoubtedly primary and often "armoured" with cancrinite where calcite abuts against nepheline.¹ The silicate minerals in carbonatites are usually patchily distributed and often the rock shows a streakiness simulating flow-structure parallel to the walls of the intrusion. Further, in some instances carbonatite plugs are composite, showing a textural variation between one member and another, in the manner of granitic complexes. Some carbonatites are demonstrably eruptive, though one hesitates to call them igneous for obvious reasons. Whether they may legitimately be regarded as magmatic depends upon how magma is defined; but it is impossible to escape from the conclusion that carbonatites which form parts of the eruptive complexes described above are magmatic in the broadest sense. By accepting the term carbonatite magma we do not imply that the latter consisted of nothing but fused carbonates; on the contrary it contained not only the components of the latter, but also of the associated silicates, as well as the volatiles responsible for the metasomatic changes. In view of these facts it is evident that one of the chief objections to the carbonatite magma hypothesis—the high melting point of the carbonate minerals—has nothing to do with the problem. It may be noted, further, that Wyllie and Tuttle² have shown by experiment that the presence of an adequate amount of water-vapour lowers the temperature at which calcite-rich liquids can occur very greatly—to the region of 650° C. even under moderate pressures. Thus the supposed physico-chemical objections to a carbonatite magma do not apply, and a reconciliation has been achieved with the field evidence.

Chemical analysis often reveals the magmatic nature of carbonatites by demonstrating relatively large amounts of such elements as barium, strontium, niobium, cerium, etc., which give rise to minerals of economic importance, such as pyrochlore, the chief source of niobium, and, on a much greater scale, of apatite. In addition to these relatively rare elements it is evident that the magma was rich in H₂O, CO₂, Ca⁺⁺ and alkalis, especially, but it was notably deficient in silica.

From the foregoing account certain important generalizations emerge. Some nepheline-syenites and rocks of ijolite type have resulted from metasomatism involving in some cases granitic rocks. Although the metasomatic aureoles frequently surround plugs of carbonatite, this is not always so, though it may be argued that,

¹ This is a useful indication of the presence, in a thin section, of nepheline which otherwise might pass unnoticed.

² Wyllie, P. J., and Tuttle, O. F., "The system CaO-CO₂-H₂O and the origin of carbonatites," *Journ. Petr.*, 1 (1960), pp. 1-46.

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although not exposed at present ground-level, it may well be hidden below. When the nephelinic rocks are associated with visible carbonatites, the latter are often seen to intrude into the former. It has been argued that as the carbonatite demonstrably post-dates the fenitization at the present level of erosion, it cannot have been the cause of the metasomatism; but evidence of this kind may all too easily be misinterpreted. At any stage before the ultimate immobilization of the magma, its uprise would have been preceded by a wave of metasomatizing emanations which fenitized the country-rock surrounding the conduit. The more slowly up-rising magma might then develop intrusive relationships towards the fenites; or alternatively, mobilized nephelinic rock-material might intrude into the material occupying the conduit. The problem is three-dimensional and the time-factor is important.

With regard to the nature of the magma responsible for these manifestations, it was evidently very different from that connected with normal volcanic activity. On general grounds it appears to have been much more basic than normal basaltic magmas: it very effectively desilicated the surrounding rocks as described above. Further, apart from the rarer elements already noted, it was rich in volatiles of which CO_2 was the most significant, and which controlled the whole sequence of events and determined the course of crystallization. The high CO_2 content resulted in the elimination of Ca'' from the silicate rocks, with the suppression of the anorthite and diopside constituents. Hence any potential plagioclase in the original magma is represented by desilicated albite—nepheline—while the associated pyroxene is generally aegirine instead of augite. The elimination of Ca'' -bearing components "boosted-up" the alkalinity of the rocks, though there is the possibility of further increase in alkalis by gaseous transfer from deeper sources.

The real nature of the magma must remain uncertain: no samples of the complete magma are available for study, and we can judge only from the effects it produced upon the contiguous country-rocks. In complexes like those at Alnö and Fen the site of the volcanic conduit is now occupied by carbonatites, as though a carbonatite magma-fraction had flushed out its forerunners, or had replaced them in some other way. There is no possible way of deciding to what extent the composition of the main stream of magma was modified by fenitizing reactions. The problem may be likened to that of deciding on the composition of liquids which had passed through the waste-pipe of a chemical laboratory, from the effects of corrosion of the pipe itself.

More direct evidence of the nature of the magmas occurring in regions where alkali-rich igneous rocks occur is forthcoming from

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the lavas themselves, of course, and this aspect of the problem is considered in the subsequent section.

(3) *Origin of the feldspathoidal lavas*

In view of the extraordinary diversity of type encountered among the lavas and dyke rocks in the central and East African Volcanic Province it is reasonable to attempt to decide how many kinds of magma were involved, and how they were interrelated. In this connection one thing must be made quite clear at the outset: every genuinely eruptive rock represents a magma-fraction; and the lavas with which we are concerned cover the widest possible range of compositions from ultrabasic to acid, but they are dominantly basic to ultrabasic.

On the basis of their composition they appear to fall into three groups. In addition to the feldspathoidal lavas with which we are immediately concerned, olivine-basalts, basalts, trachybasalts, mugearites and trachyandesites are also widely distributed and prove the availability of ordinary basaltic magmas in different stages of evolution. Secondly, basanites, basanitoids and tephrites also occur. Their significance and possible mode of origin have already been considered under "trachybasalts." Some typical members of this group are, say, 75 per cent or more basalt or olivine-basalt, and they must represent a parental basaltic magma which has not progressed far towards the ultimate alkali pole: the original composition has definitely been modified, but only to a slight extent. This conclusion leads us to the third group which includes trachytes, phonolitic trachytes and phonolites. These also have already been reviewed in an earlier chapter. Rocks of these kinds are generally accepted—on the grounds of their field relations—as the ultimate differentiation products of fractional crystallization of basaltic magma: they represent liquid fractions of extreme composition in which the maximum concentration of alkalis is achieved. Apparently in the same direct line of descent are the strongly alkaline basic to ultrabasic lavas represented by the nemafites and corresponding types. All these groups of rocks were presumably derived from basaltic magma originally; but they represent widely different stages in the evolutionary process. On the one hand unmodified basaltic magma is represented by the basalts and olivine-basalts; while on the other we have perhaps the same magma which had differentiated to the limit in the direction of increasing alkalinity, represented by the trachytes and phonolites. Varying degrees of desilication, carried to the extreme in some instances, could produce nemafite (nephelinite) and kindred rock-types from magma-fractions of this range of compositions.

This is admittedly the point of view of a magmatist, and embodies

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the simplest processes whereby this assemblage of rock-types might be produced. There is, of course, no guarantee of its correctness, and it is freely admitted that the same results might have been achieved in other, more complicated ways; but these suggestions will have served their purpose if they provide a basis for discussion.

In the foregoing account we have approached the problem from two different angles: firstly, from the interpretation of the field evidence provided by the coarse-grained rocks, with special reference to nepheline-syenites, the ijolite-melteigite series and carbonatites; and secondly from the facts of association and distribution of the corresponding lavas. It remains to correlate the two. The relationship between nephelinitic lavas on the one hand and their equivalents in the coarse-grained category on the other is one of the most elusive petrogenetic problems with which we are concerned. The case of phonolites has already been noted when these lavas were described. They are widespread, especially in the African volcanic fields, and vary little in petrographic type. The same may be said of foyaïtes among nepheline-syenites; and the two are of approximately the same composition. There can be no doubt that, in a deep-seated environment, phonolitic magma would crystallize as foyaïte; and conversely, if the corresponding volcanic phase had been preserved above known complexes containing magmatic foyaïtes, it would include phonolitic lavas. Therefore one variety of nephelinitic magma is represented by the phonolite-foyaïte association; but it is not the only one. Certain coarse-grained basic igneous rocks of essentially gabbroic composition but containing varying amounts of nepheline or the chemically allied analcite, are equivalent to the widely distributed more basic nephelinitic lavas—tephrites and basanites. The latter are as close to basalt and olivine-basalt as theralites, teschenites and essexites are to olivine-gabbro.

Rocks consisting of nepheline associated with mafic minerals but containing no plagioclase fall into another category altogether. Rocks of this composition are encountered in the coarse-grain category as urtites, ijolites and melteigites. Their fine-grained equivalents occur widely in Africa as nemaïtes (or "nephelinites" of the old terminology). Nemaïte magma exists today in the magma lake at Nyiragonga in the Congo. Crystallization at a deep-seated level would inevitably give rise to ijolite, while differentiation would produce the complementary urtites and melteigites of this well-known series, given time and a favourable environment.

The silicate portion of the whole magma (apart from its volatiles) may well be represented by ijolite; but again it would be dangerous to equate a complete magma with a single rock-type which can only represent a part of it. The complete picture is elusive. We have

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to recombine the materials in the magmatic rocks with those represented by the carbonatites and with the fugitive constituents: the product might bear some resemblance to the central and East African magma-type.¹

¹ King, B. C., and Sutherland, D. S., "Alkaline Rocks of Eastern and Southern Africa, Pt. III, Petrogenesis," *Sci. Prog.*, **48** (1960), p. 711.

This valuable paper appeared after this chapter had been written; but there is no important disparity between the writers' conclusions concerning these problems as they arise in Africa and our own, based on some personal knowledge of the Fen Complex (M.K.W.) and certain of the South African complexes (A.K.W.).

THE LAMPROPHYRES

THE lamprophyres were so named by von Gümbel (1887) when describing the "mica-traps" of the Fichtelgebirge, and had reference to the lustrous character of some types due to the presence of abundant phenocrysts of biotite. The term lamprophyre is now applied to a group of melanocratic dyke rocks irrespective of the nature of the ferromagnesian minerals of which they largely consist.

In any scheme of classification it is difficult to fit in the curious rocks grouped under this heading, for some are associated in the field, and are therefore genetically linked, with each of the major rock-groups dealt with in the previous pages. Specific types might therefore have been dealt with under the headings granitic, syenitic, dioritic or gabbroic rocks; but something is to be gained by considering them together as one group. As rock specimens, most are unconvincing. They may closely resemble normal igneous rocks in the medium and fine-grain groups, and from the examination of hand-specimens alone would almost inevitably be misidentified. Identification is possible only with knowledge of the field relations and microscopic characters.

Speaking generally the lamprophyres are strongly porphyritic with abundant phenocrysts of any of the following: dark mica, augite, hornblende and olivine. These are set in a groundmass of alkali-feldspar normally, though in one group it is plagioclase near andesine in composition, while another group comprises the feldspar-free lamprophyres.

Undoubtedly the most striking phenocrysts occur in the mica-bearing lamprophyres: the abundance and large size of the biotite crystals causes a very distinctive appearance in the hand-specimen, while under the microscope the perfect idiomorphism of the mica is unique. In common with all the other mafic constituents, the biotite may be corroded to almost any degree; while internal bleaching with the development of a complementary dark margin is a constant feature. Less commonly they may display a striking colour zoning (Fig. 121).

The amphibole in lamprophyres is of two very different types: it is a green common hornblende in lamprophyres of dioritic affinities;

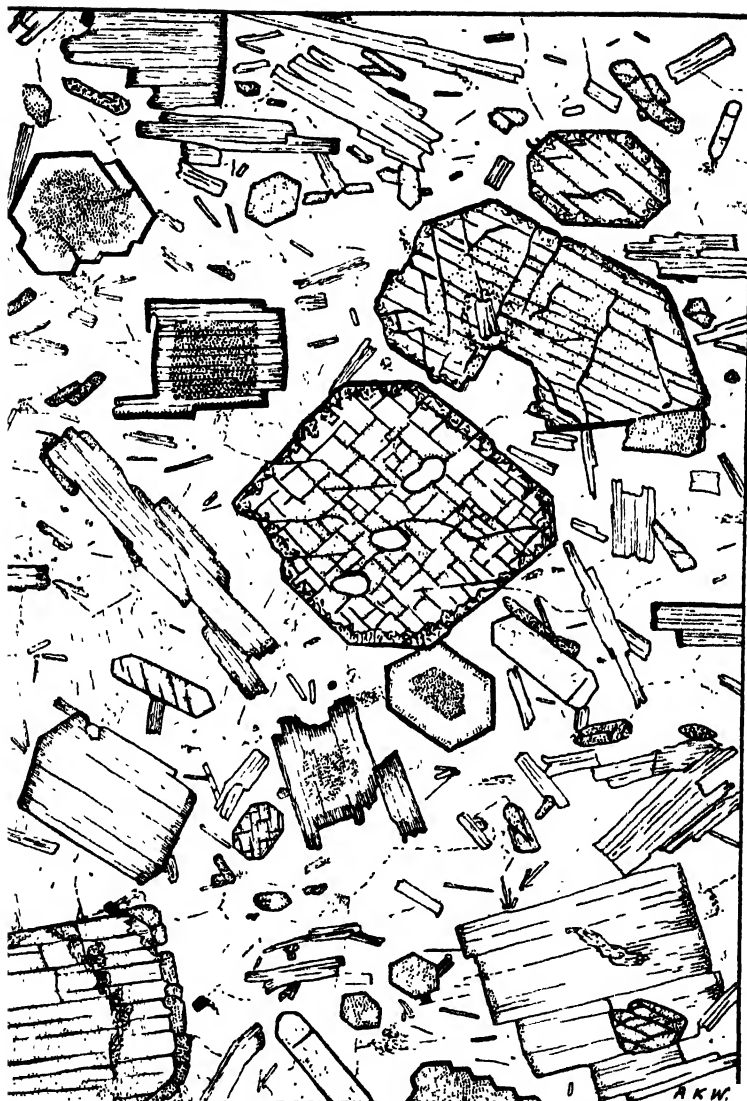


FIG. 121

Augite-minette, South Hill, St. Helier, Jersey.

Phenocrysts of diopside pyroxene subordinate to biotite, some crystals of which are beautifully zoned. Hornblende rare. The chief accessory is apatite in two generations: large euhedral phenocrysts (*e.g.* bottom, centre) and minute needles. Orthoclase forms the matrix in which all the other components are embedded; it is ideally fresh.

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but is a sodic variety, often barkevikite, in those associated with sodic syenites such as foyaite.

Similarly a light-greenish diopsidic augite is commonly seen in association with biotite in lamprophyres associated with granites and diorites; but titanaugite is often abundant in lamprophyre-dykes of foyaitic affinities. In the rock illustrated in Fig. 121 the pyroxene is zoned, a greenish core being surrounded by a colourless outer zone, while the margins are slightly granulated and in places surrounded by a reaction rim of granular bright green hornblende.

Lamprophyres of all types may contain pseudomorphs after olivine, as isolated euhedral crystals, or as synneusis aggregates. A particularly striking feature of many lamprophyres is the close association of perfectly unaltered biotite with completely pseudomorphed olivines. Although serpentine may occur in these pseudomorphs, very frequently the latter consist chiefly of carbonate. Although usually identified as calcite, it is reasonable to expect this carbonate to be magnesite, ankerite or at least dolomite. The carbonate is by no means restricted to the olivine pseudomorphs, however; it is often so abundant in the body of the rock as to cause effervescence with dilute acid. In even the freshest lamprophyres the olivine is consistently pseudomorphed; but actually these rocks are, in many instances, highly altered. In extreme cases they consist almost entirely of secondary minerals such as chlorite, carbonates, quartz, chalcedony and limonite. It is difficult to decide to what extent this is due to weathering; but there is little doubt that it is largely a late-stage, autometamorphic effect, due to a residual concentration of magmatic water, carbon-dioxide, etc.

The mafic minerals described above may occur as a second generation in the groundmass, associated in the commoner types with abundant alkali-feldspar and, in the feldspar-free group, with analcite. Occasionally the feldspar is quite fresh, water-clear in thin section, and poikilitic towards the other components. More commonly it is microlitic and may be much altered. Rarely the groundmass consists of light-brown glass which devitrification converts into fibrous subspherulitic aggregates of orthoclase microlites. Minute octahedrons of magnetite may be plentifully scattered throughout, while a high content of apatite, varying from stout prisms to delicate needles, is characteristic.

Classification and Nomenclature

Since in so many lamprophyres it is difficult or impossible to identify the feldspar accurately, while on the other hand the most striking and most easily identified components are the mafic minerals, the most useful general classification is based on the latter, thus:

mica-lamprophyre, hornblende-lamprophyre, augite-lamprophyre.

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But when the rock is sufficiently well-preserved to allow identification of the groundmass minerals, each of these main categories may be subdivided in the manner suggested by Rosenbusch, as shown in the accompanying table.

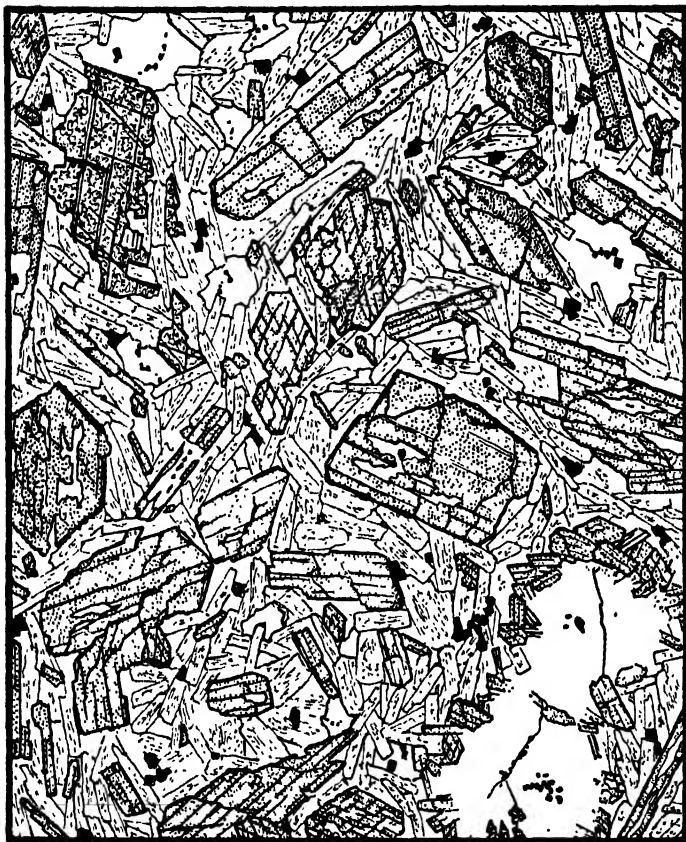
TABLE SHOWING ESSENTIAL MINERAL COMPOSITION OF
SOME TYPES OF LAMPROPHYRES

	With Orthoclase.	With Plagioclase.	No Feldspar.
Biotite	MINETTE (if with augite, augite-minette)	KERSANTITE (if with augite, augite-kersan- tite)	ALNÖITE (with melilite)
Common Hornblende	VOGESITE	SPESSARTITE MALCHITE (= aphyric Spessartite)	
Barkevikite and/or Augite		CAMPTONITE (barkevikite in type-rock from Campton Falls, New Hampshire)	MONCHIQUE (with analcite)

The advantage of Rosenbusch's classification lies in the fact that the relationship between a lamprophyre and the parental magma from which it was derived is indicated. Thus the types with dominant orthoclase—minette and vogesite (Fig. 122)—are associated with granites; the plagioclase-bearing types, kersantite and spessartite, are allied to diorites, and are often associated in the field with microdiorites. Finally, camptonite and monchiquite are associated with highly alkaline deep-seated rocks such as foyaite and other sodic syenites.

Petrographic Characteristics and Petrogenesis

The chemical characteristics of lamprophyres are shown in the table on p. 385. It will be noted that, although they contain less alkali than the aplites, in proportion to their silica percentage they are rich in alkalies, while the CaO is very high. In the mica-lamprophyres



M. K. W.

FIG. 122

Vogesite, Ards Peninsula, Ireland.

Common hornblende, stippled; magnetite in small octahedra; small areas of quartz, the largest with a rim of small hornblendes. Orthoclase chiefly in small lath-shaped crystals. Rhombohedral carbonate, top, left.

a proportion of the K_2O goes to form biotite, and in consequence of the high content of this basic silicate, such rocks with a silica percentage of under 50 may contain a good deal of free quartz.

The presence of much CO_2 and water is characteristic, and in this and other respects lamprophyres stand in contrast to aplites, with which they may be closely associated in the field, and to which they are in a sense complementary.

The strongly corroded and highly altered state of the phenocrysts in many lamprophyres proves the former to have been "out

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ANALYSES OF SOME TYPES OF LAMPROPHYRE

	I. Minette (Osann, Clarke).	II. Kersantite (Osann, Rosenbusch).	III. Vogesite (Osann).	IV. Camptonite.	V. Monchiquite (Osann).	VI. Alnöite, Alnö, Sweden.
No. of Analyses	10	20	4	15	16	
SiO ₂	49.45	50.79	52.62	46.46	45.17	27.30
Al ₂ O ₃	14.41	15.26	14.86	14.45	14.78	8.95
Fe ₂ O ₃	3.39	3.29	3.60	3.79	5.10	8.87
FeO	5.01	5.54	4.18	9.87	5.05	7.01
MgO	8.26	6.33	8.55	4.59	6.26	12.34
CaO	6.73	5.73	5.86	7.48	11.06	17.18
Na ₂ O	2.54	3.12	3.21	4.23	3.69	0.38
K ₂ O	4.69	2.79	2.83	1.96	2.73	2.99
H ₂ O	2.43	3.10	2.70	2.05	3.40	5.27
TiO ₂	1.23	1.02	0.54	4.01	1.90	3.68
CO ₂	0.61	2.61	—	—	—	2.17
Other constituents . .	1.25	0.42	1.05	0.97	0.86	4.06
	100.00	100.00	100.00	99.86	100.00	100.20

IV. Camptonite. Skaergaard, E. Greenland. E. A. Vincent, *Quar. Jour. Geol. Soc.*, 109 (1953), p. 38.
 VI. Alnöite, Alnö, Sweden (Anal. R. Blix). H. Von Eckermann, *Sver. Geol. Undersök.*, 36 (1948), p. 103.

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of sympathy with their environment," and this fact is interpreted as implying different origins for phenocrysts and groundmass. Bowen regards the phenocrysts as crystal accumulations, and stresses the alkali-rich nature of the matrix in which they are embedded. Tidmarsh, in describing some of the Exeter lavas, which show many of the characters of lamprophyres, invoked mixing and interaction between two residua: the one, a "depth residuum," highly charged with crystals of mafic minerals; the other, an alkali-line residuum, such as might, in the pure state, have crystallized as aplite.

Thus lamprophyres are abnormal rocks: they represent small highly-specialized magma-fractions, and must be carefully distinguished from dyke-rocks which have originated in other, more normal ways. There is a modern tendency, much to be deplored, to use the term "lamprophyric" as an adjective to describe any relatively melanocratic facies of syenitic or dioritic dyke-rocks. Thus hornblende in association with an "intermediate" plagioclase constitutes the normal assemblage to be expected in dyke-rocks derived from a dioritic source. Just as the parental rock-types vary widely in the proportion of light to dark constituents, so their dyke-equivalents must be expected to show similar variation. Some are melanocratic, but lack the other attributes of true lamprophyres, the corroded phenocrysts and evidence of interaction between accumulated crystals and an alkali residuum. They are, in fact, merely melanocratic microdiorites.

The lamprophyres which agree most closely with the postulated conditions of formation are the mica-lamprophyres.

In the (so-called) feldspar-free lamprophyres the **monchiquites** are outstanding. The name has reference to the Caldas de Monchique in Portugal, whence the original monchiquites were described by Rosenbusch (1890). These rocks are compact and black in hand-specimens, and superficially resemble basalts, but often contain very large phenocrysts (or xenocrysts) of bronzy biotite. Ideally they contain no feldspar, but otherwise resemble other types of lamprophyre, though probably they are even richer in dark minerals. The latter may be olivine, augite often titaniferous, amphibole or biotite. These minerals are embedded in an isotropic base, which in some cases has been identified as analcite, though in others it appears to be glass and may have the composition of melilite. In the original monchiquite the base has lately been shown to have the composition of a mixture of basic plagioclase and nepheline. Obviously the mere presence of an isotropic base cannot be considered the diagnostic feature of a rock-type, and re-examination of this group seems necessary. If the suppressed phase had been allowed to crystallize, the true affinities of the rock would have been clear. If, for example,

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it was plagioclase, then the appropriate name to apply would be **camptonite**: the term **monchiquite** should be restricted to basic lamprophyres with an analcitic base. These rocks frequently contain analcite in a more obvious form—as small circular areas resembling vesicles, but termed **ocelli**.

Monchiquites are not common rocks in Britain. Interesting examples have been described from the Orkney Islands,¹ and include a biotite-monchiquite containing many phenocrysts of olivine associated with large biotites. Augite occurs as smaller crystals making up a large part of the groundmass, while the base has been decomposed to a mixture of calcite and fibrous zeolites: its original nature remains unknown. A monchiquite dyke is intrusive into the Old Red Sandstone of Monmouthshire.² As is so often the case, this lamprophyre contains many corroded xenocrysts, including augite crystals up to 6 inches in length, plates of biotite 2 inches across and grains of quartz. The groundmass is analcitic, so the rock is a true monchiquite.

The **alnöites** are sometimes regarded as a type of feldspar-free, biotite-rich lamprophyre, allied to monchiquite. The type has already been referred to, as it contains melilite.

Examples of the feldspar-bearing lamprophyres are widespread in Britain, and a comprehensive list would occupy much space. A few typical occurrences are noted, to serve as illustrations only. The North Country lamprophyres occur as narrow dykes over an area extending from Teesdale to Furness, and from Bassenthwaite to Ingleton—a circular area with a diameter of 50 miles, centred about Shap Fell. On account of their radial disposition about the Shap granite, these lamprophyres are considered to be genetically related to it. The types represented are minettes and kersantites, though the distinction is not easily made on account of alteration of the groundmass. Mica-lamprophyres are well represented among the latest minor intrusions in the island of Jersey,³ and are noteworthy on account of the ideally fresh condition of some of them. The example illustrated in Fig. 121 is one of the most attractive. Most of the Jersey lamprophyres are narrow vertical dykes of minette, but one or two have been identified as camptonite and monchiquite, though, in the latter case, the diagnostic analcite does not seem to have been observed.

As might be expected the dyke phases connected with the Caledonian complexes of southern Scotland include hornblendic types—

¹ Flett, J. S., *Trans. Roy. Soc. Edin.*, **29** (1900), p. 387.

² Boulton, W. S., "On a Monchiquite Intrusion in the Old Red Sandstone of Monmouthshire," *Quar. Journ. Geol. Soc.*, **67** (1911), p. 460.

³ Smith, H. G., "New Lamprophyres and Monchiquites from Jersey," *Quar. Journ. Geol. Soc.*, **92** (1936), p. 365.

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vogesites and spessartites together with less common kersantites.¹ A typical example of the former is illustrated in Fig. 122, from the Ards dyke-swarm, in North-East Ireland. This rock shows a tendency for both chief minerals to be euhedral—the “pan-idomorphic texture.” The corroded quartz xenocryst surrounded by small prisms of hornblende (bottom right-hand corner) is characteristic.

¹ Phillips, W. J., “The minor intrusive suite associated with the Criffell-Dalbeattie Granodiorite Complex,” *Proc. Geol. Assoc.*, 67 (1956), p. 103.

THE ULTRAMAFITES, ANORTHOSITES AND THE PHENOMENA OF LAYERING

INCLUDED in this chapter are the descriptions of a very varied group of rock-types, most of which are represented in layered intrusions, and some of which are monomineralic. They are confined to plutonic environments and have no extrusive equivalents, so that they share a number of common problems in regard to their origin. This provides the justification for including anorthosite—a notably leucocratic rock-type—in with the various ultramafic rocks.

The account of layering phenomena may be read either in conjunction with the petrographic description of gabbroic and noritic rocks; or with reference to the bearing of layering on the origin of monomineralic rocks, so many of which are products of extreme differentiation of “gabbroic” magma.

I. THE ULTRAMAFITES

In the account of gabbros, norites and troctolites we were concerned chiefly with rocks of average composition. Those of extreme composition referred to briefly in the introductory paragraphs in that chapter remain for consideration. They may be divided into the following groups:

- (1) peridotites, in which olivine is dominant and is associated with other mafic minerals, *feldspar being excluded*;
- (2) pyroxenites, some monomineralic, others bimineralic;
- (3) picrites which contain the mineral assemblages of peridotites, but with *accessory plagioclase* in addition. They are thus the connecting links between olivine-gabbros and olivine-norites on the one hand and peridotites on the other.

One important point must be made at this juncture: we are strictly concerned with rocks which are in the direct line of descent from olivine-basalt magma. There are, however, other rocks which fall within the rock-groups named above, but which were derived from

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different sources. This fact is proved largely by their field relations, rather than by their chemical and mineral composition. In other words they belong to a different suite, the constituent rock-types in which display alkaline affinities. Thus mica-peridotite and mica-pyroxenite are to a limited extent alkaline: the mica which they contain may be regarded as the heteromorphic equivalent of olivine plus orthoclase. The latter plays no essential part in the make up of calc-alkaline pyroxenites and peridotites. Nevertheless if these two series of ultramafites were described separately considerable repetition would be unavoidable.

We therefore propose to deal with the ultramafic rocks collectively regardless of their petrogenetic affinities.

PYROXENITES

The general relationship between gabbros, norites and pyroxenites is indicated in the table on p. 307. Monomineralic varieties may consist, theoretically, of any of those clino- and ortho-pyroxenes which are relatively Mg-rich. Actually described varieties of reasonably wide distribution are bronzitites and hypersthénites.

Naturally the reader will expect "augitite" to occur in this list; but this name is not available as it was long ago applied to a type of feldspar-free lava containing prominent augite phenocrysts. **Diallagite** is fully described by its name. Of the pyroxenites the recently described monomineralic **aegirinite** from the Kola Peninsula has the distinction of being the only rock of its kind of distinctly alkaline (sodic) type.

Pyroxenites are quantitatively important rocks in some of the great stratified lopoliths, particularly the Bushveld and Stillwater Complexes and the Great Dyke of Rhodesia. In the first named, hypersthene- and bronzite-pyroxenites, some of them virtually monomineralic **bronzitites**, form sheets of great extent: even when only a few feet thick, a bronzitite scarp-feature in the Bushveld may be visible extending along the strike for many miles until it ultimately passes over the horizon. One such pyroxenite contains the famous "Merensky Reef" which yields platinum in exploitable amounts. In the Stillwater Complex a thickness of 2,500 feet consists of bronzite-pyroxenite almost identical with the Bushveld occurrences, associated at the base of the lopolith with bronzite-peridotite ("harzburgite"). In both occurrences chromite is a constant accessory, and layers of chromitite (described below) are exploitable commercially.

The most striking accessory in some varieties of bronzite-pyroxenite is the bright emerald-green clinopyroxene, usually termed chrome-diopside, though chrome-augite is, in most cases,

THE ULTRAMAFITES, ANORTHOSITES AND LAYERING

a truer name. These rocks are two-pyroxene pyroxenites and the most satisfactory way of naming them is to combine both mineral names, using the dominant one for the rock-type, and the minor one as the qualifier. Thus in **augite-hypersthene**, augite is second in importance to dominant hypersthene. Another type contains both diallage and hypersthene in unspecified proportions; and according to the relative amounts it is feasible to distinguish between **diallage-hypersthene** and **hypersthene-diallagite**—both terms are self-explanatory, and in our opinion superior to “websterite” (G. H. Williams, 1890) the name chosen for this particular pyroxenite from a locality in North Carolina, and, to the confusion of students, still in use.

Mica-pyroxenites are chemically not very different from some kinds of mica-peridotites: the latter are slightly more under-saturated, but both types are relatively rich in K' and Al'' .

That these rocks lie beneath some of the existing lava fields is proved by their occurrence as xenoliths in the lavas themselves which include such distinctive types as the leucite- and kalsilite-bearing types in the (former) Belgian Congo and in Uganda. Augite-peridotite and even olivine also occur in this association, so that the alkaline affinities of the latter are established in these particular cases. Biotite-pyroxenite is represented among the ejected blocks of deep-seated rocks for which Monte Somma is well known. In Northern Ireland a member of the Newry Complex has the composition: biotite 45, augite 27, hornblende 20, actinolite 2, iron-ores 3 and apatite 3. A rock of this mineral composition is difficult to name. In the description of the complex¹ the name chosen was “biotite-pyroxenite,” but as there is nearly twice as much mica as pyroxene, perhaps “augite-hornblende-biotite” gives a truer indication of the mineral composition of the rock.

PERIDOTITES

The essential features of peridotites as noted above, are their exclusively ultramafic character—no plagioclase is allowable—and the dominant role of olivine. This is the sense in which the term peridotite has been used since redefinition by Rosenbusch in 1877. One type is monomineralic and consists of olivine as the only essential constituent. It is normally not olivine-rock in the sense of consisting of the pure mineral; a constant accessory is chromite, and it is only reasonable to permit a very small quantity (say up to 5 per cent) of other accessories, without the necessity of changing

¹ Reynolds, D. L., “The eastern end of the Newry Complex,” *Quar. Journ. Geol. Soc.*, **90** (1934), p. 585.

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the name. The generic name for these rocks is *olivinite*,¹ while such specific names as forsteritite, hortonolitite etc. may be used as appropriate. Instead of *olivinite*, the less satisfactory name *dunite* (Hochstetter, 1859) is still often used. The type takes its name from a New Zealand locality, Mount Dun, and consists of olivine with chromite as an accessory. Olivinites of this type have been recorded as one of the minor rock-types occurring in layered basic complexes, in which the occurrences vary from sheets a few feet in thickness to films only a few crystals thick.

All olivinites do not occur in the form of sheets in layered complexes, however. Some occur as pipes and dykes rising from the upper surface of larger olivinite sheets, while in the Bushveld Complex carrot-shaped bodies cross-cutting the layered members of the Complex present an interesting problem. The detailed forms of these cone-olivinites have been very thoroughly studied in the course of mining operations for platinum. Obviously cone-olivinites must have originated under conditions significantly different from those obtaining during the formation of sheet-olivinites. The latter, by general consent, represent accumulations of olivine crystals which settled out of a body of magma, under gravity control; but the material forming cone-olivinites must either have displaced or replaced its own volume of the layered rocks which it cuts. From the textural angle there is nothing in these rocks to indicate that they were formed otherwise than by direct crystallization *in situ* from a "melt" of olivinitic composition (see Figs. 123 and 124). On the other hand Hess describes as "*secondary dunites*" certain irregular bodies occurring in the Stillwater Complex, embedded in, and formed from, serpentinite, by a process of reconstruction *in situ*.² Bowen and Tuttle have shown experimentally that such reconstruction can be effected at temperatures as low as 500° C., under pneumatolytic conditions.³ The Stillwater "*secondary dunites*" are somewhat richer in iron and coarser in grain than the sheet-dunites, and are stated to be almost identical with the hortonolitite-cones of the Bushveld. There we leave the matter for the moment: it is certainly a thought-provoking problem.

In chemical composition olivinites show analyses like those of the individual minerals. With the incoming of a little alumina, this combines with MgO or FeO to give the almost constant accessory, spinel. With rather more silica, orthopyroxene appears.

¹ Unfortunately olivinite is very differently defined by Johannsen (Vol. IV, 402). He uses this name for "dunites carrying a considerable amount of pyroxene, or amphibole, or even biotite." In our classification such rocks are typical *peridotites*.

² Hess, H. H., "The Stillwater Igneous Complex, Montana," *Geol. Soc. Amer.*, Mem. 80 (1960), p. 62.

³ Bowen, N. L., and Tuttle, O. F., "The system MgO-SiO₂-H₂O," *Amer. Journ. Sci.*, 29 (1935), pp. 157-217.

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Orthopyroxene-peridotites. In view of the chemical relationship between olivines and the corresponding orthopyroxenes, it will be realized that, with a slightly less degree of under-saturation, monomineralic olivinite will give place to enstatite-peridotite, bronzite-peridotite or hypersthene-peridotite, according to the Mg : Fe ratio



FIG. 123

Olivinite ("Hortonolite"), Mooihoek, Bushveld Complex, South Africa. Anhedral grains of hortonolite showing cleavages and dendritic plates of iron ore regularly orientated in two directions.

of the rock concerned. In most occurrences (especially the Bushveld and Stillwater Complexes)¹ the orthopyroxene falls within the bronzite range. Although no actual figures are available, it is probable that **bronzite-peridotite** is the most abundant type of peridotite, not only in the two complexes named above, but in all parts of the world. Unfortunately all peridotites are prone to alteration into "serpentine," and in many occurrences little trace of the original minerals is left: this is the case, for example, in the Lizard Complex in Cornwall; but little difficulty is experienced in inferring from the structure of the serpentinite that the "bastite-serpentine" was originally bronzite-peridotite. Ideally fresh rock of this type occurs in a very different setting elsewhere. In the Kimberley

¹ These bronzite-peridotites are the "harzburgites" in H. H. Hess's account.

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district in South Africa deeply eroded volcanic pipes filled with kimberlite (described below) yield an extraordinary number and variety of blocks of ultramafites, some of very large size. Among them are bronzite-peridotites, quite unaltered, consisting of dominant olivine in closely packed crystal grains showing peripheral

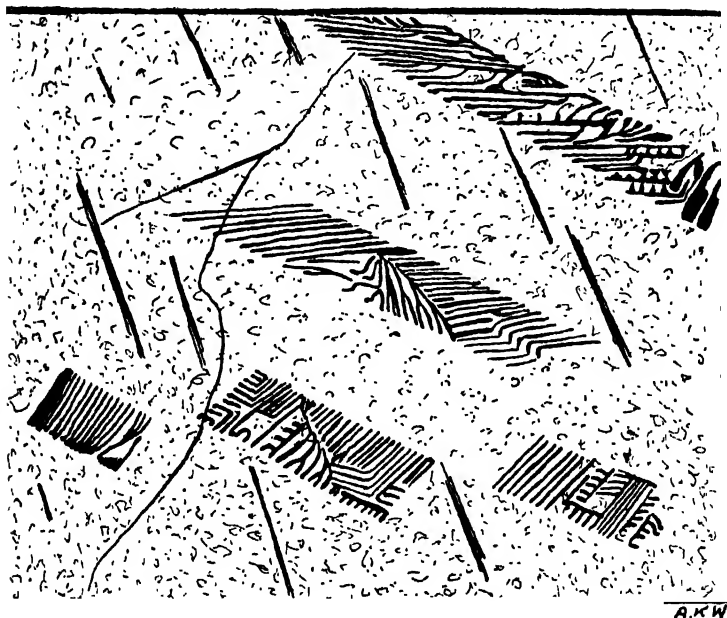


FIG. 124

A.K.W.

Regularly orientated plates of dendritic iron-ore in hortonolite. Same rock as that illustrated in Fig. 123, drawn under $\frac{1}{4}$ -inch objective.

granulation due to movement during accumulation of the crystal mush. Embedded in the olivine are scattered "phenocrysts" of orthopyroxene, somewhat rounded, while translucent spinel is a constant accessory.

Not uncommonly a second pyroxene, in an accessory role, may accompany the hypersthene or bronzite. It may be augite or the much rarer chrome-augite which plays the same part in the corresponding pyroxenites, as noted above. Examples are to be found among the blocks occurring in the kimberlite pipes; but of greater significance are so-called olivine-nodules which occur in basaltic lavas in many parts of the world.¹ Specimens are to be found in

¹ Obviously the term olivine-nodules as applied to these bodies is a misnomer. A nodule is a concretionary body formed *in situ* in various kinds of rocks. They have nothing in common with the peridotitic xenoliths occurring in basaltic lavas and dykes.

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most teaching collections, often as mineral specimens of olivine; but even with a hand-lens, and more clearly under the microscope it is seen that these are rock-fragments, *i.e.* xenoliths of peridotite of a distinctive kind. They consist of a coarse-grained aggregate of olivine, enstatite, chrome-augite and spinel—a typical peridotitic mineral assemblage. They often show clear evidence of selective magmatic corrosion.

Augite-peridotites do not appear to be widely distributed, but they are represented in the Garabal Hill–Glen Fyne Complex.¹ Augite is the most abundant mineral, occurring as large irregular crystal grains poikilitically enclosing numbers of olivines and associated with small amounts of both hornblende and dark mica. The addition of plagioclase to this assemblage gives augite-picrite, a much more widely distributed rock-type.

Hornblende-peridotite might well figure in a list of possible types of peridotite, though it is indeed difficult to see how a rock consisting exclusively of olivine and hornblende could be formed, in view of the wide separation of these two minerals in the discontinuous reaction series. However, hornblende does occur, with other mafic minerals in some rare types of peridotite, including parts of the Cortland Complex in New York State.² Shand has pointed out that there is continuous variation in several directions, and in a rock-body of such variable composition, it is unprofitable to attempt restricting the definition of **cortlandtite**: it is a hornblendic hypersthene-peridotite, containing in addition to the three chief components some malacolite (a clinopyroxene colourless in thin section) and mica. It is reminiscent of scyelite, described below.

Mica-peridotites. Of the peridotites those which carry abundant mica are evidently as strongly alkaline as any peridotites can be. After olivine, mica (typically a strongly phlogopitic variety), is the most abundant mineral and accounts for the relatively high content of K' and Al''' . The most widely distributed and the most important of the mica-peridotites is, beyond doubt, **kimberlite**, which takes its name from the diamond-mining area in the Transvaal, where it occurs in a series of deeply eroded volcanic pipes. Unfortunately the kimberlite itself is very thoroughly brecciated and "altered" and as the "blue ground" of the miners, it does not lend itself to detailed petrographic study. Actually the kimberlite pipes are of far greater interest on account of an extraordinarily varied assortment of xenoliths, large and small, of ultramafic

¹ Nockolds, S. R., "The Garabal Hill–Glen Fyne Igneous Complex," *Quar. Journ. Geol. Soc.*, 96 (1940), p. 451.

² Shand, S. J., "Phase Petrology of the Cortlandtite Complex," *Bull. Geol. Soc. Amer.*, 53 (1942), p. 409.

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rock-types, including eclogites, which occur in the pipes and represent material which has been brought up from great depths. Kimberlite pipes penetrate sedimentary rocks of Karroo age in Tanganyika and are cut by dykes of the same type. The pipe-filling material, the dykes and country-rock are heavily carbonated. The kimberlite, as is commonly the case, is extremely altered to a mixture of chlorite, talc and carbonates; but occasionally it is solid enough to analyse and some of the olivine is fresh. Outstanding chemical features are the extraordinarily low SiO_2 percentage (25 to 30), high magnesia (24 to 25), TiO_2 (3 to 4), and CO_2 (about 10 per cent). Diamond-bearing kimberlites occur also in Equatorial and West Africa and in eastern Siberia (Yakutsk), where again they are noteworthy for the ultramafic xenoliths which they contain, and rarely in the eastern Canadian Shield, where the pipes penetrate through the Keewatin lavas. Diamonds occur rarely in the nearby Drift. Carbonate-bearing kimberlite dykes have been discovered around some of the carbonatite ring-complexes, for example, at Alnö.

Among peridotites of more mixed composition scyelite (Judd, 1885) named from Loch Scye in Sutherlandshire is a phlogopite-hornblende-peridotite. In hand-specimens scyelite is a very distinctive, heavy, dark-coloured rock, with large (one inch) mica and hornblende crystals containing black olivines. The texture is poilitic and the consequent lustre-mottling is most striking. The features seen in thin section are illustrated in Fig. 125.

Certain Scottish peridotites with alkaline affinities form the basal portion of the well-known differentiated Lugar Sill from Ayrshire. In addition to dominant olivine this rock contains both titanaugite and barkevikite, sometimes the former, sometimes the latter being in excess. Barkevikite is, of course, one of the NaFe-rich amphiboles; but the alkaline character of the rock is more clearly demonstrated by the fact that the complex, of which it forms a part, consists largely of teschenite (analcite-bearing) and theralite (nepheline-bearing). The Lugar peridotite grades upwards into picrite, containing the same mineral assemblage, but with labradorite in addition.

PICRITES

The most important diagnostic feature of picrite (Tschermak, 1866) is the presence of subordinate plagioclase; otherwise picrites are very close to peridotites. The introduction of feldspar into a peridotitic mineral assemblage involves crossing a phase boundary. In principle it is unsound to define a given rock-type in such a way that some specimens fall in one compositional field, but others in an adjoining one. Therefore there can be no latitude so far as the occurrence of plagioclase in peridotites is concerned. At the point

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where the bronzite-olivine assemblage gives place to bronzite, olivine and *plagioclase*, we pass out of the peridotite, into the picrite, field. The point may be profitably illustrated by an actual example. In the basal portion of the Stillwater Complex, among the dominant bronzite-peridotites are variants carrying olivine and

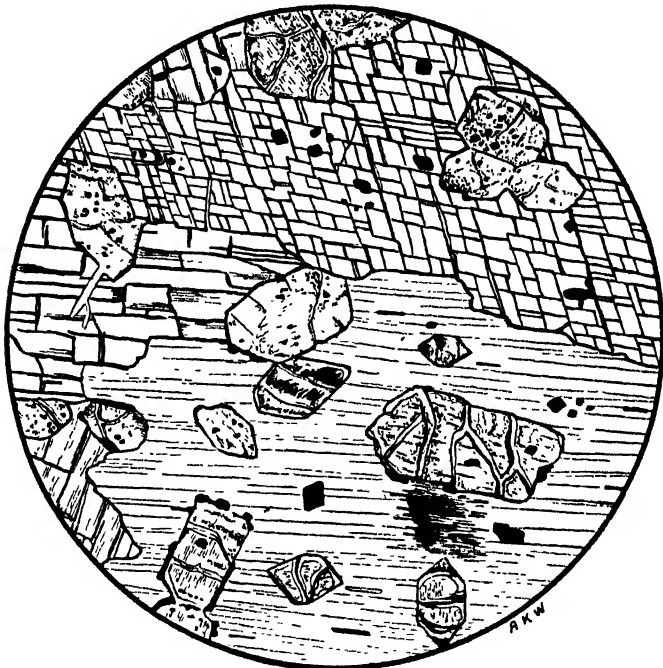


FIG. 125

Scyelite, Loch Scye, Sutherland.

Upper part of the field occupied by basal and vertical sections of nearly colourless amphibole; the lower part by phlogopitic mica. Serpentine pseudomorphs enclosed poikilitically in both minerals. Magnetite in octahedra. Hexagonal plates, seen end-on, causing schiller structure in the mica (bottom, right).

bronzite (two-thirds of the whole rock) with chrome-augite and plagioclase accounting for the rest. The amount of plagioclase is not stated; but that is really beside the point. *Some* plagioclase occurs and therefore the assemblage is picritic, not peridotitic according to definition—our definition admittedly. The rock-name applicable is “chrome-bearing bronzite-picrite.”

Variants of this type are distinguished as **enstatite-picrite** or **hypersthene-picrite** according to the composition of the pyroxene.

A well-known picrite, mineralogically much like the Lugar picrite

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referred to above, forms part of the small island, Inchcolm, in the Firth of Forth. This rock contains a large assortment of minerals and is therefore much used for teaching purposes. Olivine, partly altered into a vivid yellow serpentine, is abundant, and is accompanied by titanite, in reaction relationship with subordinate barrovite, and abundant phlogopitic mica, now largely represented by bluish green chlorite. The felsic minerals in the Inchcolm picrite are difficult to identify. The amount of basic plagioclase is, in any case, small, and it is further obscured by analcization. Even the analcite is turbid and usually has to be "taken on trust."

Calc-alkaline hornblende-picrites, containing common hornblende as the chief mafic component additional to olivine, occur, for example, in the Plymouth district in Cornwall, Anglesey in North Wales and Colonsay and Glen Orchy in Scotland.

Although there are excellent grounds for precisely defining the boundary between peridotites and picrites, there are no similar reasons for limiting the amount of plagioclase acceptable in a picrite before it grades into melagabbro. The dividing line is arbitrary. This point may be illustrated by reference to the 500 ft. layered sill which forms the Shiant Isles off the Scottish coast.¹ In the lowest visible portion a so-called picrite occurs consisting of olivine (59), augite (10), plagioclase (26), iron-ores (2) and zeolites (3 per cent). With over a quarter of the whole rock consisting of plagioclase and with a colour index of only 71, we think the rock would be more appropriately named melagabbro.

Before leaving the picrites it will be well to remember that they are the only ultramafic coarse-grained rocks which have their equivalents among the lavas, namely the picrite-basalts including both olivine-rich (oceanites) and augite-rich varieties (ankaramites).

MICAITES (GLIMMERITES)

Under these synonymous terms we group certain rocks which are no less monomineralic than olivinite, or pyroxenite, though they are unrelated to the suites considered above. The name for a rock consisting essentially of dark mica is difficult to choose: micaite is the obvious choice and is in use, but some people prefer the Germanic form, *glimmerite*, which is certainly more euphonious. The specific terms *biotitite* and *phlogopitite* are self-explanatory but must be used only if the mica is of the appropriate composition: a "biotitite consisting of phlogopite" has been recorded as quoted, but is certainly not accurately named!

How these mica-rocks originate is problematical. They are certainly of deep-seated origin, and as they do not occur in igneous

¹ Walker, F., "Geology of the Shiant Isles," *Quar. Journ. Geol. Soc.*, 86 (1930), p. 355.

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layered complexes it is possible that they should be regarded as metamorphic rather than igneous rocks. They are represented among the ultramafic xenoliths in the kimberlite pipes in the Kimberley district, and occur also among the granulites and gneisses forming the Basement Complex in southern Nyasaland. There is no rock in the fine-grain category which closely matches these coarse-grained rocks, though some minettes (mica-lamprophyres) are exceptionally rich in biotite, to the extent of up to 80 per cent.

CHROMITITES

The amount of chromium calculated as Cr_2O_3 in undifferentiated noritic or gabbroic magma is only a fraction of 1 per cent, but in a small specimen from a chromitite layer the figure may rise to a high value. Apparently the magma quickly becomes effectively supersaturated with Cr_2O_3 and a swarm of small octahedra is precipitated. These crystals filter down through the magma and accumulate on a "floor" where further movement is impeded by crystallization of silicates. The interprecipitate magma at this early stage must be capable of precipitating pyroxene and plagioclase, and in the case illustrated both these minerals are seen acting as hosts to the chromite octahedra (Fig. 126). Had the magma remained fluid for a longer time segregation would have been more complete, the chromites would have been more closely packed, while the interprecipitate material would have been proportionally less. It will be noted that where the concentration of chromite is highest, the crystals are largest and so are those representing the interprecipitate material: a single plagioclase crystal grain spans the field at the bottom of the diagram.

Another Bushveld type contains the same high concentration of chromites, associated in this case with dull greenish bronzite crystals in a whitish plagioclase matrix. This rock has been enriched in two components, firstly in chromite, and secondly in bronzite—both presumably by gravity differentiation. The rock is chromite-enriched melanorite, complementary in composition to the first type. A third type known to the writers is virtually a two-mineral rock, and consists of a coarsely crystalline three-dimensional "mosaic" of large irregular bronzites, of pegmatitic dimensions—up to half an inch or more in diameter—in which the closely packed chromites are embedded poikilitically, so that hand-specimens display a distinctive lustre-mottling. The rock is in effect a chromite-enriched bronzitite, whose formation may be explained as follows. A few small bronzites were enmeshed in the chromite crystal concentrate. As no other silicate minerals were present at this level, these seed crystals gave bronzite an advantage over the other silicates; a diffusion-gradient was set up and as a consequence, Mg- and Fe-ions



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FIG. 126

Contact-zone between leuconorite and chromitite, Dwars River, Bushveld Complex, Transvaal, South Africa.

The outlines and nature of the twinning in the plagioclase are indicated. Hyperssthene stippled; chromite in single crystals and groups of octahedra increasing in size and concentration downwards.

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migrated towards the rapidly enlarging bronzites, while ions which could not be incorporated migrated in the opposite direction—chiefly upwards. From the large size of the bronzite crystals a high concentration of volatiles, favourable to pegmatitic crystallization, may safely be inferred. The rock illustrated in Fig. 126 presents some features of interest especially in the field as seen at Dwars River in the Bushveld. The chromitite layers are more sharply defined than is suggested in the figure; they are parallel to the general stratification of the Complex and are interlayered with leuconorite. Further, individual sheets change their horizons suddenly and bifurcate in a manner closely simulating a series of intrusive sills. The field relations are extremely well displayed, and it is difficult to avoid the conclusion that here a chromite concentration was sufficiently lubricated by the interprecipitate material as to develop an intrusive relationship towards “rafts” of plagioclase crystals with which it became associated before finally freezing-in.

MAGMATIC IRON-ORE ROCKS

Magnetite and ilmenite display many features similar to those described above for chromite in their occurrence in layered intrusions. Thus magnetite and ilmenite both occur in small quantities in gabbroic and noritic rocks and their derivatives, and, like chromite, may become concentrated into layers which consist almost exclusively of ore-minerals, either ilmenite or magnetite separately, or more commonly of the two in close association. There is, however, one significant difference between natural concentrates of chromite on the one hand and of iron-ores on the other. The former is probably always an early precipitate, while magnetite commonly forms late in the crystallization sequence. Further, iron-enrichment is proved by the changes in composition of successive crops of olivine and pyroxene crystals precipitated at successive stages in the formation of a complex such as that of Skaergaard (in which this phenomenon has been particularly studied). It is underlined, so to speak, by the evident iron-enrichment¹ in the interstitial glassy residuum in certain olivine-basalts as already described. The potentiality of the formation of a magma-fraction notably enriched in iron should be borne in mind; this may well be important in connection with the mode of origin of some of the layered iron-ore rocks.

But whatever the composition of the melt, some of the iron must become oxidized from the ferrous to the ferric state, before magnetite can be precipitated. The state of oxidation depends largely on the concentration of water-vapour in the melt.¹ If the latter is “dry”

¹ Kennedy, G. C., “Equilibrium between volatiles and iron-oxides in igneous rocks,” *Amer. Journ. Sci.*, **246** (1948), p. 529.

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a large proportion of the silicates must crystallize from it before the water content of the residuum reaches a value sufficiently high for magnetite to crystallize. This naturally affects the textural relationships between the iron-ores and the silicates. Frequently in gabbroic rocks ilmenomagnetite forms poikilitic or interstitial crystal growths which are evidently of late origin. It follows, further, that the iron ores will be more closely associated with the later, rather than the earlier-formed silicates. This means that magnetite-ilmenite layers in the Bushveld and Freetown Complexes (to quote two examples known to the writers) are closely associated with either anorthositic or pegmatitic facies of the rocks concerned. Some of the largest ilmenite deposits occur in anorthosites, as for example, in the Adirondacks. It will be remembered that chromite concentrations are normally associated with early-formed members of a complex, consisting largely of olivine and pyroxenes.

Probably the most fascinating examples of titano-magnetite rocks occur at the aptly-named locality, Magnet Heights, in Sekukuniland in the Transvaal. The iron-ore rocks are interstratified with anorthosite and norite at a high level in the Main Zone of the Bushveld Complex. The contact between an underlying white anorthosite and the overlying black iron-ore rock is sharp and striking; but although the lowest parts of the latter are almost pure, strongly magnetic iron ore, plagioclase crystals appear in the ore, at first widely scattered, but increasing in numbers upwards. They are strongly fluxioned, lying with their pinacoidal faces parallel to the top and bottom of the layer. Fuller description of these puzzling rocks and a discussion of their origin would take far more space than can be spared; but mention should be made of "reaction rims" which separate iron ore from the plagioclases and involve granular Fe-rich olivine, shells of cummingtonitic amphibole and Fe-rich biotite as well as symplectic pyroxene-plagioclase intergrowths (growing like myrmekite) shown in Fig. 127. The writers cannot believe that these magnetite rocks represent gravity-sorted concentrates of iron-ore crystals which accumulated *in situ*: the movements which presumably brought the different layers into juxtaposition would account for the fluxioning of the plagioclases embedded in the ore. The phenomena would be most easily explained if the original magma-fraction was itself an iron-rich residuum. This possibility is supported by the fact that above the magnetite occurs a **ferro-anorthosite** consisting of bytownite (An_{63}) 87, iron-ore 11, pyroxene 1, and lepidomelane 1 per cent. Without the iron-ore the rock would be an anorthosite of the same composition as that occurring below the magnetite band. This rock has a normal igneous texture; the iron-ore is subophitic and is largely later than the dominant plagioclase. It clearly represents a late magma fraction and illustrates

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FIG. 127

Reaction Phenomena and Symplectic Structure in Magnetite-Anorthosite, Magnet Heights, Sekukuniland, South Africa.

In this section titano-magnetite runs between plagioclase crystals, while a distinctive feature is the reaction rim of fayalitic olivine which insulates the iron-ore from the feldspar. Myrmekite-like symplectites of orthopyroxene and plagioclase are lobed into the labradorite. Mg-rich mica is intimately associated with the olivine at one point, while the latter contains dendritic magnetite.

Orthopyroxene, light stipple; fayalitic olivine, close stipple; mica, lined; iron-ore, opaque. Twinning indicated in the plagioclase.

the close association of anorthosite and iron-ores which we inferred above, from theoretical considerations. In this rock we have potentially both rock-types—anorthosite and magnetite rock—present in a homogeneous layer: all we need is a process for the separation of the liquid from the crystalline phase; and in that connection we can only suggest filter-press action at the appropriate moment in the evolution of the complex.

ECLOGITES

Eclogites are of special interest and present problems of interpretation. The name is an old one, introduced in 1822 by Haüy for rocks occurring in the Fichtelgebirge, and which still remain the most typical eclogites. Eclogite is of notably high specific gravity;

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and consists of bright red garnets set in a bright green crystalline matrix. Two components are essential in eclogites—red garnet of the pyrope-almandine type, and the bright green pyroxene, omphacite. Accessories sometimes present include brilliant emerald-green chrome-diopside and, strangely enough, quartz and kyanite. It has been shown by analysis that eclogite has much the same chemical composition as olivine-gabbro, but contains none of the normal gabbroic minerals: plagioclase has been completely eliminated and the other components have been broken down and redistributed. Labradorite has been split, as it were, into albite and anorthite “minerals”; the former has been desilicated into jadeite, which is incorporated in the pyroxene, omphacite. The silica released may appear as free quartz in spite of the basicity of the rock. The anorthite molecule goes into the garnet which, though fundamentally a type rich in magnesium derived from olivine and original pyroxene, does contain a certain proportion of the “lime-garnet” molecule ($\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$). The kyanite represents excess alumina and silica not required in the adjustments mentioned.

Typical eclogites are well represented among the blocks in the “blue ground” which fills the deeply eroded vents in the Kimberley region in South Africa. In Britain eclogites occur in northern Scotland in the so-called Lewisian inliers near Glenelg—the first record in this country (J. J. Teall, 1891)—in central Sutherland and central Ross-shire. They there form small outcrops over a wide area, and are closely associated with injection gneisses which were originally basic igneous rocks impregnated by pegmatite. True eclogite forms only a small part of the eclogitic bands which contain cores of the typical red-garnet, omphacite rock. The latter grades into related types in which feldspar, quartz and green hornblende become important. These changes are regarded as due to retrogressive metamorphism, the ultimate product of which is garnet-amphibolite, an undoubted metamorphic rock.

The chief problem of the eclogites concerns their origin. They are believed to be products of “plutonometamorphism”—to lie on the borderline between the igneous and metamorphic. It is uncertain, however, whether they represent basic (gabbroic) igneous rocks which have been metamorphosed under extremely deep-seated conditions, or whether they represent basic *magma* which crystallized under such conditions. In either case, they are of outstanding interest and provide such a useful study of heteromorphism that they cannot well be omitted.

SERPENTINITES

Although as a general rule special names are not considered necessary for “altered rocks,” serpentinites are the exception to

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the rule. These consist essentially of mixtures of the different serpentine minerals, chrysotile (fibrous), antigorite (platy), lizardite (massive) and bastite in varying proportions. Popularly these rocks are often named "serpentines"; but it is awkward and in a sense incorrect to use the same name for a mineral and for the rocks composed largely of that mineral: therefore we use "serpentinite" for the rocks and "serpentine" for the mineral.

Serpentinites are compact, variously coloured rocks, often banded, blotched and streaked with bright green and red, the general body of the rock being any colour between light greyish-green and greenish-black. On account of their attractive colouring and because they can be shaped and polished easily, they are used for ornamental purposes.

In some specimens relics of original minerals may remain and give an indication of the nature of the unaltered rock. From such evidence it is clear that some serpentinites correspond in composition with, and were presumably produced from, ultramafites of various types. In some nothing but olivine (and chromite) can be inferred from the existing texture, and the term "dunite-serpentinite" is appropriate. In others bastite pseudomorphs after bronzite or other orthopyroxenes are prominent and are embedded in other forms of serpentine which clearly indicate that they are pseudomorphous after olivine, and the rock was evidently a bronzite-peridotite. It may safely be inferred, therefore, that in general serpentinites represent ultramafites, dominantly peridotites of various kinds, also picrites and pyroxenites which have been profoundly altered, the changes being of a simple type, involving only hydration of the original silicates. Sometimes the alteration has been so complete that there is no indication that the minerals of which the rocks are composed are other than in their original condition: in other words some serpentinites *look* as if they were intruded as serpentinite magma from which the constituent serpentine minerals separated as primary constituents.

With regard to distribution, serpentinites occur in two very different associations: (1) in the basal parts of layered complexes, as, for example, in the Ultrabasic Zone of the Stillwater Complex; and (2) as sheet-like bodies or lenses occurring in orogenic zones, and frequently bounded by thrust-faults. Outstanding examples of the latter are afforded by the Great Serpentinite Belt in New South Wales, and the comparable belt in southern New Zealand, where individual masses may extend for 400 miles and may be on average 4 miles thick. They are almost vertical and deeply dissected, and the wholesale serpentinization is a most impressive feature. These sheets are in part peridotites but largely serpentinites. Similarly serpentinites forming "belts," "lenses" or "masses" extend throughout

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the central tectonic zone in Japan. They are of special interest in that they are associated with dykes of two complementary groups: (1) leucocratic—comprising albitite, quartz-albitite, anorthosite, and trondjemite; and (2) melanocratic—including hornblendite, titanaugite and some gabbroic types.

In Britain serpentinites are represented on a modest scale in the Shetland Islands (Unst), the Outer Hebrides, at Portsoy in Banff, in the Mona Complex in Anglesey and, on the most extensive scale, in the Lizard Complex thrust over the Devonian rocks in southern Cornwall.

With regard to origin the problem is similar to that of the spilites: serpentinites are altered rocks, but where, when and how the alteration was effected are matters of uncertainty.¹ Without the water they contain, and allowed to crystallize normally, serpentinites in the main would be converted into peridotites: the "secondary dunites" of the Stillwater Complex are believed to have originated in this way. In dealing with orogenic, as distinct from layered-complex, serpentinites we are concerned with peridotites *not* associated (in the same rock-bodies) with other rock-types from which they might have been derived by gravity-controlled crystal accumulation. We are led to infer, therefore, that the orogenic serpentinites were derived from primarily intrusive peridotite. The physical state of the latter at the time of its intrusion provides one of the major problems of petrogenesis on which a considerable amount of light has been thrown by an investigation of the system $\text{MgO-SiO}_2\text{-H}_2\text{O}$.² This was the first study of its kind involving high water-vapour pressures, equivalent to a depth of 6 miles below the surface and temperatures up to $1,000^\circ\text{C}$. Even under these conditions the silicates forsterite and enstatite remained completely crystalline. Therefore water-vapour alone does not lower the melting points sufficiently to make a peridotite magma of this composition a reasonable proposition, and makes it almost certain that orogenic serpentinites were moved and intruded under compressional conditions as lubricated crystal mush. This is consistent with the almost negligible metamorphic effects observable around these bodies. Further, the experiments showed that crystalline olivine remains unaltered in the presence of any quantity of water down to temperatures of 500°C . This is, therefore, the highest temperature at which serpentinization can take place.

It is widely believed that serpentinites are connected in a significant fashion with spilites. Certainly both now occur associated

¹ Hess, H. H., "Serpentinites, Orogeny and Epirogeny" in "Crust of the Earth" Symposium, *Geol. Soc. Amer. Special Paper*, 62 (1955).

² Bowen, N. L., and Tuttle, O. F., "The system $\text{MgO-SiO}_2\text{-H}_2\text{O}$," *Bull. Geol. Soc. Amer.*, 60 (1949), pp. 439-60.

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with rocks which accumulated in a geosynclinal environment. Up to a point they share certain chemical characteristics, *i.e.* high water and CO₂ content, but they are widely separated in time: spilites are generated in the manner described above during an early phase in the development of a geosyncline; but serpentinites are associated with the orogenic phase following the collapse of the geosyncline; for example in this country the well-known Cambro-Silurian geosyncline dates from the opening of the Cambrian period; the earliest spilitic eruptions occur in the Arenig Stage of the succeeding Ordovician period, and the main period of folding coinciding with emplacement of serpentinites followed at the end of the Silurian.

II. ANORTHOSITES

As already noted anorthosites are pure or nearly pure plagioclase rocks: in a sense they are plagioclases, consisting usually of labradorite or bytownite, but the composition of the feldspar depends to some extent upon the nature and environment of the anorthosite. Plagioclase from sheets low down in a layered sequence tends to be more calcic than that from higher levels, while andesine, indicating dioritic rather than gabbroic affinities, is widespread in some of the large plutonic anorthosite bodies. Although some anorthosites are very light coloured rocks, in keeping with their low colour index, others are quite dark as the feldspar is schillerized, as in the well-known rock from Newfoundland which is so much used for demonstrating this property, while some of the Norwegian anorthosites are purplish brown. According to the definition adopted here, up to 10 per cent of minerals other than plagioclase may occur: with more than this amount anorthosite grades into leuconorite, leucogabbro or leucotroctolite according to the dominant mafic mineral present. Actually these mafic minerals are often so widely scattered that it is difficult to estimate proportions.¹ The nature of the coloured silicate tends to vary with the composition of the plagioclase: thus olivine accompanies bytownite, augite and/or bronzite tend to accompany labradorite, while hornblende may occur in the dioritic anorthosites containing andesine. Magnetite and ilmenite, usually intergrown, are constant accessories, and are sometimes concentrated into thin layers and segregations in exploitable amounts. Many anorthosites display nodular weathered surfaces due to widely spaced masses of pyroxene in ophitic relationship to the feldspars. The "mottled anorthosites" of A. L. Hall from the Bushveld are of this type.

Anorthosites are found in two strikingly different environments:

¹ Some authors apply the term anorthosite to rocks containing more than this amount of coloured and accessory minerals: thus the Stillwater anorthosites contain 85 to 100 per cent plagioclase.

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firstly they are widespread in layered basic complexes, occurring in sheets varying in thickness from mere streaks to layers many yards thick; and secondly and on a much more important scale they form major rock-bodies of batholythic dimensions in Pre-Cambrian shield areas, where they may outcrop over vast areas. Thus in the Canadian Shield the Saguenay anorthosite occupies approximately 5,800 square miles of surface area. These rock-bodies ensure that anorthosite is by far the most important quantitatively of all the monomineralic and near monomineralic rocks.

With regard to origin, much that has been said concerning the ultramafic rocks including olivinites and pyroxenites applies also to anorthosites, especially those occurring in layered complexes. Sheet-anorthosites to a large extent must represent layers of sorted plagioclase crystals: this is clear from their field relations with other gabbroic or noritic derivatives. This does not imply, of course, that the crystals sank through the magma from which they were precipitated: the specific gravity of these basic plagioclases is not very different from that of ordinary basic magma, so that conceivably the crystals might be buoyed up forming a "plagioclase raft" or if they came under the influence of rising convection currents they might form a roof accumulation. The general picture involves a layer of plagioclase crystals out of which any mafic silicates had been separated by gravity, while interstitial magma might rise or be squeezed out.

The stage of crystallization with which we are concerned is important: much MgFe would have been extracted from the magma by the crystallization of olivine and much of the pyroxene. Thus what is left must have become enriched in plagioclase-building components and also in volatiles. Yoder¹ has shown experimentally that the addition of H₂O to melts of diopside and plagioclase has a profound effect upon the eutectic relationship between the two minerals, lowering the freezing temperature as might be expected, and changing the composition of the lowest freezing point mixture (the eutectic) far towards enrichment in anorthite. In general terms, a differentiating gabbroic magma containing the ingredients of plagioclase and pyroxene has its composition changed by increasing concentration of water, becoming progressively more feldspathic.²

¹ Yoder, H., "Synthetic Basalt," *Ann. Rep. Dir. Geophys. Lab., Carnegie Inst., Washington* (1953-54), pp. 106-7.

² Experimental data for the melting temperatures of anorthite and diopside in a "dry" system are shown diagrammatically in Fig. 68. The eutectic composition is 60 per cent diopside and 40 per cent anorthite, and the temperature is about 1,270° C. With sufficient water, the temperature is lowered to below 1,100° C. and the eutectic composition shifted to only 30 per cent diopside and 70 per cent anorthite. The effect of other components such as would occur in nature are unknown; but there can be little doubt that they would further lower the freezing points, bringing these well within the range of natural magmas.

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Now the actual magma with which we are concerned is interstitial to a plagioclase crystal concentrate; its composition will almost certainly lie well within the plagioclase field and therefore the first phase precipitated from it must be plagioclase, and with free diffusion a pure anorthosite will be formed. Even without postulating free diffusion and thus conveniently dispersing the unwanted components, anorthosite within the terms of the definition used in this book would be formed from a 60 per cent plagioclase crystal concentrate provided that the interstitial liquid contained 70 per cent of plagioclase substance: the anorthosite would be 90 per cent pure, the remainder would be pyroxene and iron ore. Provided that the pyroxene crystallized from the interstitial liquid and not from suspended nuclei, it would form widely dispersed, large poikilitic masses of the kind already noted: there would be no suspended seed crystals which would act as nuclei for outgrowth, and therefore the resulting mode of crystallization is in complete contrast with that of the plagioclase. These large, irregular ophitic pyroxenes may thus enclose large numbers of plagioclase crystals. By contrast the plagioclase precipitated from the interstitial liquid forms either outgrowths from the original crystals or a second generation, easily recognized in thin sections. These textural features are very distinctive.

The explanation summarized above has been applied to the Freetown Complex, Sierra Leone,¹ where the conditions for the formation of a residual feldspathic melt appear to have been particularly favourable. The anorthositic layers are underlain by troctolites, troctolitic gabbros and other olivine-rich rocks. The downward filtration of early precipitated olivines is clearly demonstrated by the field-relations, as indicated in Fig. 128, and it can be seen that the anorthosite layers represent complementary accumulations of precipitated plagioclases. The textural features seen in thin sections demonstrate the two stages of crystallization noted above. The question now arises as to whether the same processes may reasonably be expected to have operated in the case of the great batholythic anorthosites of the Pre-Cambrian shield areas. It is tempting to argue that as these rocks are petrographically identical with sheet-anorthosites in layered complexes, they must have been formed in the same way; but this is not necessarily so. Naturally it is necessary to examine the possibility of these rocks having been formed by direct crystallization from a melt of their own composition. This has already been ruled out as impossible for other monomineralic rocks on account of the abnormally high temperatures

¹ Wells, M. K., and Baker, C. O., "The anorthosites in the Colony Complex near Freetown, Sierra Leone," *Col. Geol. and Min. Resources*, 6, No. 2 (1956), pp. 137-58.

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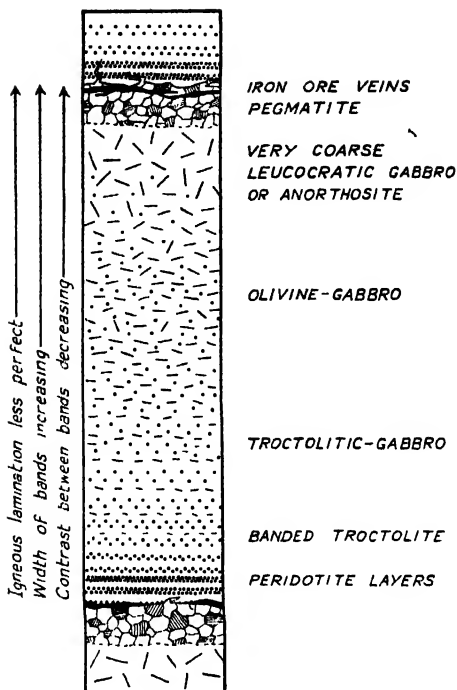


FIG. 128

Idealized layered unit of the Freetown intrusion. Individual units range from 200 to 500 feet in thickness.

involved. The same objection applies to anorthosites as far as dry melts are concerned. However, Yoder's experiments have shown that with adequate water, magmas of leucogabbro composition (*i.e.*, approaching that of anorthosite) may form. Since such magmas would involve high water-vapour pressures they could only be formed under plutonic conditions. It is significant that no anorthositic lavas occur. Although magma approaching an anorthositic composition is likely to be formed under certain conditions, this does not constitute an entirely satisfactory explanation of the batholithic anorthosites of the Pre-Cambrian, because in many instances they are associated with only very limited metamorphic effects. It still seems necessary to postulate some mechanism of crystal-accumulation as in the case of the layered anorthosites. Gravity sorting of the components is a possibility, though an argument against this is the apparent absence of the enormous amounts

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of basic and complementary ultramafic rocks which the hypothesis demands: they are not associated in anything like the requisite amounts, so far as can be determined. True, the argument is not wholly conclusive because practically nothing is known as to the depths to which anorthosite extends underground, nor what underlies it. Nevertheless, it is generally agreed that by some means the latter has become separated from its former associates, presumably by movement in the form of a lubricated crystal-mush during compressional earth movements. During such movement some granulation of a cleavable mineral seems inevitable: it is significant therefore that peripheral granulation is often seen in thin sections, indeed in many instances a pseudoporphyrity texture has resulted, with large, somewhat rounded plagioclases being embedded in a "groundmass" consisting of granulated material. The texture is cataclastic. The necessary deformation and squeezing of largely crystalline material is possible under orogenic and plutonic conditions though, of course, this has obviously not been a factor in the formation of the layered anorthosites described above.

To prove in any specific case whether or not such movement of a crystal mush has taken place involves detailed study of the structural relationships within the rock-body and its surroundings. Such studies have been carried out in the Adirondacks, for example.¹

III. LAYERING PHENOMENA

The most striking feature of the great gabbroic and noritic complexes is their composite character: they are built up of a number, often a large number, of rock-types. In some cases it has proved possible to establish the fact that differentiation by fractional crystallization has affected magma of "average" basaltic composition after injection of the latter. In other cases it appears likely that successive injection of magma has been followed by differentiation in place. As a consequence the major units become split into smaller ones forming layers, seen as bands of different composition, texture or colour on a rock-face. Consequently such terms as "banded gabbros" have become commonplace since first used by Alfred Harker when describing the igneous rocks of Skye.² "Band-ing" is not a wholly appropriate term to apply to a three-dimensional structure—"layering" is preferable.

On the grandest scale layered units are recognized in the great

¹ Balk, R., "Structural survey of the Adirondack anorthosite," *Journ. Geol.*, **38** (1930), pp. 289-302; and Buddington, A. F., "Adirondack Igneous Rocks and their Metamorphism," *Geol. Soc. Amer. Mem.*, **7** (1939).

² "Tertiary Igneous Rocks of Skye," *Mem. Geol. Surv.* (1904). Apart from a full description of the banded gabbros, the illustrations are most useful and convincing.

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basic complexes enabling the whole to be divided into "zones" such as the Critical Zone of the Bushveld Complex¹ and the basal Ultrabasic Zone of the Stillwater Complex.² These are of the order of several thousand feet in thickness. Each such zone is capable of subdivision into thinner lithologically distinctive units which to a varying degree are repetitive. Thus in the Freetown Complex, Sierra Leone, troctolitic gabbros alternate with anorthosites (Fig. 128); while in the Stillwater Complex the latter alternate with gabbros, the individual rock-types forming layers some hundreds of feet thick. Not all major layering of this kind is repetitive: individual distinctive layers of extraordinarily uniform character and thickness may be of very wide extent and must underlie large areas, measurable in hundreds of square miles. The Merensky Reef in the Bushveld and the Chromitite Layer in the Stillwater are notable examples.

So far we have been concerned with major units down to layers of recognized well-known rock-types; but the layering to which the term is commonly applied is on a smaller scale and involves varying proportions of the gabbroic minerals, olivine, pyroxene, plagioclase and accessories, which may occur singly in monomineralic layers, or in close association. The thicknesses of individual layers vary from a fraction of an inch to perhaps about a foot. There may be mere streaks of chromitite in thin layers of anorthosite or olivinite; or the layering may take the form of manifold repetitions of half-inch layers of light-coloured plagioclase-rich gabbroic rock (leucogabbro) alternating with correspondingly pyroxene-rich melagabbro. These thin layers are complementary in respect of their mineral composition and hundreds of such units may be visible in a single exposure.

In small-scale layering when it is only the proportions of the component minerals which vary it is appropriate to speak of the variation as involving **rhythmic mineral layering** when it is of the kind 1-2-3: 1-2-3, etc. A gradation may be seen from a clearly defined dark base enriched in, say, olivine or pyroxene, passing up gradually into a leucocratic feldspathic top, as in the thinner differentiated layers in Fig. 129. This is comparable with grading in sedimentary rocks and suggests by analogy that these layers were built up from the base by crystal accumulation.³

The problem of accounting for layering phenomena is not a simple one. The scale, details and kinds of variation are widely

¹ Hall, A. L., "The Bushveld Igneous Complex of the Central Transvaal," *Geol. Surv. S. Africa.*, Mem. 28 (1932), p. 560.

² Hess, H. H., "The Stillwater Igneous Complex, Montana," *Geol. Soc. Amer.*, Mem. 80 (1960), p. 230.

³ Wager, L. R., "Layered Intrusions," *Meddel. Dansk. Geol. Foren.*, 12 (1953), pp. 335-49.

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different and no one process can be applicable to all cases. Successive injections of magma-fractions of contrasted composition has frequently been suggested as the probable cause of some kinds of layering. The acceptability of the process depends upon the compositions of the magma-fractions involved: in some cases it is manifestly impossible. Some of the layers are monomineralic, and it is generally agreed that melts having the composition of pure



FIG. 129

Gravity differentiated layers separated by layers of average rock in the Skaergaard Complex, in eastern Greenland.

(After Wager and Deer, *op. cit.*, plate 8.)

plagioclase, olivine, pyroxene or chromite do not exist: the temperatures involved are the melting points of these minerals and are several hundreds of degrees higher than known magmatic temperatures: olivine, for example, melts at about 1,700 to 1,800° C. This conclusion is confirmed by the absence of lavas of similar composition; while the metamorphic effects produced on adjacent rocks are negligible.

A process which can be seen to have operated in many cases is **crystal accumulation**. There can be no doubt that, in general, in sheet-like intrusions including lopoliths, crystallization commences in the coolest part, close under the roof, but accumulation of the precipitated crystals takes place on the then-existing floor, where they are spread out to form a layer. The effectiveness of the process depends upon the difference in specific gravity of the precipitated

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crystals and of the magma from which they have separated: it is most marked in the case of the accessories, including chromite, and among the silicates, olivine, which separates first and at highest temperatures. Further, there may be a definite interval between the separation of olivine and the commencement of pyroxene- or plagioclase-crystallization. This gives olivine a good start over the other minerals named, and other things being equal, should give a relatively sharp, clear-cut surface of separation in the case of olivinite (dunite) layers. This process makes no great demand on the imagination to be visualized in operation. But actually it is not quite so simple as that. The crystals sink through magma into liquid which is displaced by the growing accumulation, but never completely: some interstitial magma remains and acts as a lubricant, so that before it solidifies the crystals may be subject to disturbances like those which affect sedimentary materials before their induration on the sea-floor. Consequently structures akin to those covered by the term sedimentation structures may be developed, and include slump structures due to changes in the slope of the floor perhaps. The layer may be uparched or down-warped; it may show false-bedding, washouts and channels. In fact the aping of sedimentary structures is so close that if it were not for independent evidence proving a magmatic origin, it might well be thought that the whole complex is a sort of pseudomorph after a bedded sequence of sedimentary rocks.

A given monomineralic layer therefore represents an accumulation at a particular level of a concentration of crystals, all of one kind. It resulted from **crystal sorting**, the all-important process operative in such cases. It is not merely a matter of a slight difference in specific gravity between a particular kind of crystal and the magma from which it was precipitated: several minerals may well be involved over the same period of time. They will certainly sink at different rates due to differences of two kinds—in specific gravity, and, less obviously, in crystal habit, which will affect the buoyancy of the crystals, as will the viscosity of the magma. Convection also plays its part, which is believed in certain circumstances to be a major one. “Convective overturn” is the term used for the process whereby a roof-accumulation of crystals, after being buoyed up for a time, is perhaps quite suddenly carried down to the floor. There it may still be under the influence of convection currents which may spread and winnow the crop of crystals, for in a major magma chamber convection currents move down, along and up, so that in some places these currents will co-operate with gravity, but elsewhere the two agents of crystal-sorting will work in opposite directions, so that certain crystals may be prevented from sinking or may even be carried upwards to form a roof accumulation.

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Wager and Deer¹ have sought to explain the magnificent display of rhythmic layering in the Skaergaard intrusion, East Greenland, by sudden changes in the velocity of convection currents: rapid flow causes turbulence and maintains crystals of all the minerals uniformly in suspension, while slower movements allow the denser minerals to settle. Evidence from the Freetown intrusion suggests a process of mechanical sorting by laminar flow which passed into shearing movements as the magma crystallized. In most layered intrusions effects of flow movements are indicated by the preferred orientation of crystals, particularly of tabular plagioclases which lie with their most prominent (010) faces parallel to the layer surfaces. Fluxion is the general name given to this structure; but if the evidence points to an origin by the settling of crystals uninfluenced by lateral flow movements (as in normal sedimentation), then the structure may be termed "igneous lamination" (Wager and Deer, *op. cit.*).

Nearly all the accepted explanations of rhythmic layering invoke mechanical principles of crystal sorting; but we should not leave the subject without stating our belief that chemical processes, such as some kind of pulsatory crystallization, may ultimately be found to provide more convincing explanations of some of the puzzling phenomena involved,² particularly when the layers are thin and two minerals only are involved.

One other point must be made: it is relatively easy to suggest a reasonable explanation of the mineral variation observed in *one* layered unit; but the elusive feature is the rhythmic *repetition* of the particular set of conditions involved (whether purely mechanical, physico-chemical or both), on perhaps a hundred separate occasions. Whether the repetition resulted from periodically applied external stimuli, or the repeated build-up, during protracted crystallization, of conditions favouring the precipitation of different minerals in a definite sequence, remains problematical.

Cryptic layering, aptly so termed as it is not visible to the naked eye, involves systematic changes in the compositions of the individual minerals throughout a layered series. The minerals are precipitated from successively lower-temperature fractions of the magma as the layered series is built up, so that there is a tendency for lower-temperature members of each of the isomorphous mineral groups—plagioclases, olivines, orthopyroxenes and clinopyroxenes—to appear in the higher layers of the series. In other words, one

¹ Wager, L. R., and Deer, W. A., "Geological investigations in East Greenland. Part III. The petrology of the Skaergaard intrusion, Kangerdlugsuaq, East Greenland," *Meddel. om Grønland*, Bd. 105, No. 4 (1939).

² Cf. Wager, L. R., "Differing powers of crystal nucleation as a factor producing diversity in layered igneous intrusions," *Geol. Mag.*, 96 (1959), pp. 75-80.

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can observe in a layered intrusion a wide range of products of differentiation by fractional crystallization, all laid out as a continuous series in vertical sequence upwards. The most notable example is provided by the Skaergaard layered series, in which the changes of composition within each mineral group follow almost perfectly the trends indicated by the experimental researches of Bowen and his colleagues. As the plagioclase becomes progressively more sodic in the ascending sequence, so the ferromagnesian minerals show iron enrichment. The changes shown by the more important mineral groups involved in the Skaergaard Complex are summarized in tabular form below:

<i>Height above base of layered series (metres)</i>	<i>Plagio- clase</i>	<i>Olivine</i>	<i>Ortho- pyroxene</i>	<i>Clino- pyroxene</i>
2,400	An ₃₀	Fa ₉₆		Wo ₃₀ En ₂ Fs ₆₈
1,800	An ₄₀	Fa ₆₀	En ₂₆	Wo ₃₄ En ₃₂ Fs ₃₄
1,100	An ₄₅		En ₃₉	Wo ₃₂ En ₃₂ Fs ₃₆
600	An ₅₆	Fa ₃₇		Wo ₄₂ En ₄₀ Fs ₁₈
Early (marginal) phases	An ₆₅	Fa ₁₉		Wo ₄₁ En ₄₂ Fs ₁₇

Changes of the same kind, though not necessarily of the same range, characterize most major basic complexes; but in some no cryptic layering can be discerned, while in others it is repeated. In the latter case it appears likely that the complex was built up of a number of successive influxes of magma, each separate injection becoming in effect a unit of cryptic layering.

A special feature of cryptic layering has been termed **phase layering** by Hess, and involves the sudden appearance or disappearance of a particular mineral in the rock sequence, reflecting the fact, of course, that the composition of the magma had at that time passed the appropriate phase boundary. Thus in the Ultra-basic Zone of the Stillwater Complex the pyroxene is almost exclusively bronzite: precipitated augite first appears at a level of 5,600 feet above the base and represents the important stage in the evolution of the complex when, for the first time, the CaAl concentration was such that augite instead of bronzite could crystallize.

ORIGIN OF MONOMINERALIC ROCKS

As already noted, among the rocks forming layered complexes monomineralic types are encountered, and the first important step in the process of formation has already been described: it involves

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the accumulation of an essentially monomineralic layer of crystals, resulting normally from differential crystal sorting.

The second stage in the formation of a monomineralic rock involves the crystallization of the interstitial (or interprecipitate) magma. It is impossible to pack crystals so tightly as to exclude completely the liquid magma into which they sank. The composition of the latter must vary with the passage of time according to the minerals which have been separated from it; but at an early stage it could not be very different from the original magma, and therefore potentially any of the minerals appropriate to it could crystallize from the interstitial liquid trapped in the interstices in the crystal concentrate. To deal with the problem more explicitly we will consider the formation of olivinite. Olivine is precipitated so early and at so high a temperature that a layer of self-supporting olivine crystals near the floor is a foregone conclusion.¹ Further, the interstitial magma at this early stage and at this level would not be appreciably different from the original magma before intrusion. Let it be supposed that this magma is noritic. It is therefore potentially capable of crystallizing as a "mixture" of olivine, pyroxene (say bronzite), and plagioclase.

Hence selective crystallization must operate, allowing the precipitation of further olivine, while preventing the formation of bronzite and plagioclase. In the case of a pure olivine concentrate, the crystals would present a relatively enormous surface area against the enveloping liquid, and this would give olivine an advantage over pyroxene and plagioclase as these would have to set up independent nuclei as a prelude to crystallization. Therefore the settled olivines grow by enlargement, the outgrowths being in optical continuity with the original crystals. The impetus given to the crystallization of olivine might be such as to set up a diffusion gradient which would ensure the flow of the necessary MgFe-ions from the adjacent overlying magma to the growing crystals, and in the ideal case this might continue until the interstices were completely filled, and olivinite would result.

In some cases careful examination of thin sections of monomineralic rocks shows slight, but perceptible, differences between the original crystals of the concentrate and the secondary outgrowths; but this is not invariably so, and there may be no visual evidence of the two stages of crystal growth.

There are two stages, therefore, in the formation of a monomineralic rock: firstly, the accumulation of a layer of crystals of one kind, and secondly selective crystallization under conditions involving free communication with the main magma body, thus

¹ The actual floor is in most cases a thin chilled phase often of relatively fine-grain.

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ensuring free diffusion. The necessity for free diffusion was first suggested by Hess; though the hypothesis as stated here was elaborated to account for the occurrence of thin olivinite layers in the Freetown Complex,¹ Sierra Leone.

¹ Wells, M. K., and Baker, C. O., *op. cit.* See also: Wager, L. R., Brown, G. M., and Wadsworth, W. J., "Types of Igneous Cumulates," *Journ. Petr.*, **1** (1960), pp. 73-85.

THE PYROCLASTIC ROCKS

THE term pyroclastic implies fragmentary material resulting from volcanic activity of explosive type. The material forming pyroclastic accumulations is of the most varied character: it may include, or consist of, volcanic bombs or lapilli of lava (plastic when erupted), or of any type of sedimentary, metamorphic or igneous rock lying beneath the volcano, in fragments varying in size from the smallest dust particles to large blocks some feet in diameter.

Classification of pyroclasts is difficult as there are several variables including: (1) the composition of the component fragments; (2) their size; (3) their lithic character; and (4) their mode of origin. For the purposes of mere record (1) and (2) are readily combined and such terms as basaltic (or andesitic or rhyolitic) agglomerate (or tuff) are self-explanatory. Actually most pyroclasts are Acid to Intermediate in composition as a consequence of the relatively higher viscosity of magmas of these types as compared with Basic (basaltic) ones.

With regard to variation in grain-size it would simplify matters to adopt the scheme of classification used in sedimentary petrology, distinguishing as far as possible between deposits consisting of blocks (boulders), cobbles, pebbles, coarse, medium and fine sands, silt grains and clay particles in order of decreasing grain-size. In the coarse-grained category volcanic breccias are analogous with, and may closely resemble such sedimentary breccias as those of Permian age well exposed along parts of the eastern Devon coast. Cobbles and the larger pebbles compare in size (and often in shape) with volcanic bombs and lapilli. The general name **tuff** is applied to medium or fine-grained pyroclastic rocks, while "ash" is the term loosely applied to such deposits in their unindurated condition. If precision in nomenclature is desired, then a particular tuff may be described as being of coarse, medium or fine-sand grade. The silt and clay grades are also represented by volcanic dust particles which are prone to wind-transportation over long distances before deposition.

The variation in grain-size may be correlated with distance from the eruptive centre, the coarsest and heaviest material being deposited

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first and close to the volcano, while at the other extreme volcanic dust may be carried high into the air and may ultimately settle thousands of miles away. Fragments of pumice may be floated by currents for long distances until they ultimately become water-logged and, on sinking, contribute to the material forming the deep-sea deposits.

To cover variation in lithology Pirsson¹ introduced three useful terms: (a) vitric tuffs (or ashes), composed of glass fragments; (b) lithic tuffs or ashes, composed of rock fragments; and (c) crystal tuffs, composed of crystals, whole or fragmentary. It is only rarely that a particular tuff consists wholly of one of these classes of material: usually it is an admixture and may be very heterogeneous.

Pyroclasts composed dominantly of angular fragments of large size are volcanic breccias. The term "**agglomerate**" is practically synonymous and is widely used without restriction as to grain size. Normally the angular rock-fragments of which they are composed are widely variable in size—they are completely unsorted—and in the rock-types represented. Agglomerates occur in many volcanic successions, particularly at or near the base of the pile of inter-bedded lavas and pyroclasts. These are distinguished as bedded agglomerates. Lithologically identical rocks occur within circumscribed, roughly circular outcrops, and in three dimensions must have an approximately cylindrical form. These are distinguished as vent-agglomerates, and are believed to represent infillings of volcanic vents. **Explosion breccias** are special types, of greater interest than their name suggests.² They consist of blocks of country rock (sedimentary, metamorphic, or igneous) varying in size from an inch to several feet in diameter, embedded in a matrix consisting of finely comminuted rock- or mineral-fragments. In the marginal parts of the rock-body the blocks are sharply angular and scarcely displaced; but away from the contacts they become more widely separated, the amount of matrix increases and they lose their angularity, the edges and corners becoming rounded.³ It is suggested that, after the initial fragmentation, gas streamed through for a long period, carrying with it, in suspension, mineral fragments which in these circumstances could actively abrade the larger blocks. This mechanism was first suggested by D. L. Reynolds⁴ who termed it "fluidization," by analogy with an industrial process involving a

¹ *Amer. Journ. Sci.*, **40** (1915), p. 193.

² Hughes, C. J., "The Southern Mountains igneous complex, Rhum," *Quar. Journ. Geol. Soc.*, **116** (1960), p. 111, with discussion.

³ By definition breccias consist of *angular* fragments; but the most distinctive feature of these explosion-breccias is *rounding* of the included blocks. In spite of this anomaly there is no other term available, and it is appropriate when the whole rock-body is included under the term.

⁴ Reynolds, D. L., "Fluidization as a geological process . . .," *Amer. Journ. Sci.*, **252** (1954), p. 577.

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suspension of small solid particles in a gas stream. It may well be that close examination of the agglomerates filling certain volcanic vents, for example those forming such a distinctive feature of the geology of the Fifeshire coast in Scotland, may reveal rounding of the included blocks and other signs of a similar mode of origin.

Intrusive tuffs have recently been described from Tertiary¹ ring-complex fault zones within which they can be seen to intrude, and even "finger" into country-rock. Although the material consists largely of rounded fragments of rock-forming minerals including quartz, feldspars, augite, etc., of about 0.1 mm. diameter, larger rock fragments also contribute, together with blocks comparable in size with those occurring in explosion breccias, but fewer in number. The conditions of formation are believed to have been essentially the same as for the explosion breccias; but the explosive fragmentation, followed by gas-streaming, occurred at greater depths and under higher containing pressures. The gas operating in the fluidization process is believed to be supercritical water-vapour.

The phenomena described above are of particular interest and importance as providing examples of intrusion of non-magmatic (non-liquid) material. Prior to the demonstration of the effects of gas-streaming and fluidization, intrusive contacts were regarded as providing infallible evidence of the magmatic nature of the intrusive material. To this extent the problem has been complicated, particularly in the case of fine-grained rocks of granitic composition, and it is essential to examine the evidence of each case critically. It is particularly unfortunate that late-stage recrystallization affecting rocks of this composition—whatever their mode of origin—may give rise to an uninformative felsitic texture. A valuable clue may be provided by the virtual absence of metamorphic effects adjacent to an Acid rock which originated as an intrusive tuff, because of the dissipation of thermal energy in doing the work of fluidization.

Crystal tuffs are relatively uncommon and demand rather special conditions of formation. They represent phenocrysts brought up from depth by magma which must have been notably rich in crystals of the kinds which occur in the tuffs. An explosion in the vent would disrupt the column of magma and the crystals would be separated from the liquid through friction with the air. In this way accumulations of loose crystals, often of perfect form, may be built up on the flanks of the volcano, providing a profitable hunting ground for mineral collectors. Leucite and augite crystals around Vesuvius provide a striking example of the process. Not all crystal accumulations have originated in this way, of course; the olivine sands occurring locally on the Hawaiian beaches consist

¹ Hughes, C. J., *op. cit.*, p. 120.

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of perfectly euhedral crystals of olivine resulting from the rapid disintegration of olivine-rich basalts which have been subjected to active marine erosion. In view of the prevalence of feldspar phenocrysts in several widely distributed types of lava, it might be anticipated that feldspar-crystal tuffs would, among rocks formed in this way, be relatively common. They have been recorded, for example, among the Ordovician volcanics of the Lake District and North Wales, and these serve to illustrate a difficulty encountered when volcanic ashes of any kind are deposited under water in a marine environment. In such circumstances the "ashes" show normal sedimentary structures—stratification, current bedding, ripple marking and grading, together with possible admixture with normal sediment appropriate to the environment. Thus feldspar-crystal tuffs may closely resemble, grade into, and therefore be difficult to distinguish from, feldspar sands.

Reverting to the coarse-grained pyroclasts, J. F. N. Green, when studying the Borrowdale Volcanic Series in the Lake District, drew attention to the fact that rocks resembling volcanic breccias may result from autobrecciation (flow-brecciation), involving the breaking up of the crust of a lava flow followed by the incorporation, within the still fluid portion, of the solid rock fragments. The latter, on exposure to weathering, become etched out (or in, as the case may be, depending upon whether the blocks prove to be more or less durable than the matrix). Superficial examination of such rocks during a field survey might lead to their misidentification as agglomerates; but the study of thin sections should show that, in the case of an autobrecciated lava, both included blocks and matrix consist of lava, of the same type, though differing in minor points of detail.

The difficulty of distinguishing between lavas and tuffs is acute in the case of certain widespread rocks termed either ignimbrites or **welded tuffs**. Certainly two, possibly three different modes of origin are involved, but the products have several points in common, and on the evidence available long after the event, it is difficult, if not impossible to decide which of the alternatives applies to a particular occurrence. The main difficulty is to decide whether the rocks in question are lava or tuff. It is significant that Russian petrologists have introduced a term translated by S. Tomkeieff as "**tuff-lavas**," to cover these problematical rocks. If they are lavas, obviously the term welded tuff should not be applied. These rocks are to a large extent fragmental; but the fragmentation has resulted from two different processes. In the first instance a rhyolitic lava column in a vent was disrupted by vent explosions which formed a nuée ardente of incandescent lava fragments lubricated by steam and erupted with explosive violence. This was the mode of origin that

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Marshall¹ had in mind when he coined the word **ignimbrite**; and seems consistent with contemporary accounts of the "burning cloud" eruption of Mount Pelée in Martinique in 1902. Such explosively disrupted molten rhyolite, dacite or even andesite is undoubtedly pyroclastic by definition. The glass shards on settling retained sufficient heat and plasticity to be flattened and welded together and this is the chief diagnostic feature of true welded tuffs.²

In the second case, however, explosive volcanic activity played no part, except to the extent of clearing the way for the quiet upwelling of (usually) rhyolitic magma in a state of effervescence, being continuously expanded by escaping volcanic gases, and issuing under low-pressure conditions from fissures, forming a "glowing avalanche" of foaming lava. This material is indubitably **lava** and to call it "welded tuff" would create an utterly false impression of its mode of origin. Some fragmentation is inevitable in the conditions obtaining during the eruption, but was incidental to the foaming process: it did not result from explosive volcanic activity.

The district around Katmai and including the famous Valley of Ten Thousand Smokes has come to be regarded as the type locality for rocks of the kind under discussion, and the products of the 1912 eruption provide the standard for comparison.³ Following a recent re-examination of the type area Howel Williams describes the material as massive unstratified "glowing avalanche deposits . . . of dust-size to bomb-size fragments of extremely vesicular white rhyolite pumice." He emphasizes that "the magma frothed quietly to the surface instead of being propelled high above the vent."⁴

Products of Katmaian eruptions, sometimes identified as welded tuffs, elsewhere as ignimbrites, are of very wide distribution: in fact one gets the impression that this is the characteristic style of eruption wherever magma of the appropriate degree of viscosity and gas content is, or has been, available. Among "fossil" examples many rocks, formerly regarded as normal, massive rhyolites, have on re-examination been found to be "Katmaian" (ignimbritic). Typically these rocks occur as extensive, flat-topped rock-bodies, without obvious connection with a vent, resting on a basal glass-dust tuff. Welding has occurred extensively, though in ancient

¹ Marshall, P., "Acid Rocks of the Taupo-Rotorua Volcanic District," *Trans. and Proc., Roy. Soc. New Zealand*, **64** (1935), pp. 323-66.

² Mansfield, G. R., and Ross, C. S., "Welded rhyolitic tuffs in South Eastern Idaho," *Trans. Amer. Geophys. Union*, **16** (1935), pp. 308-21.

³ Williams, Howel, "Problems and Progress in Volcanology," *Quar. Journ. Geol. Soc.*, **109** (1954), pp. 311-32.

⁴ Actually two magmas, one rhyolitic, the other andesitic, were in eruption simultaneously, so that dark andesitic scoriae are mixed in amongst the light rhyolitic material; but this is not an essential feature of eruptions of Katmaian type.

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examples the evidence may have been largely destroyed by deuteric modifications including devitrification and spherulitic and axiolitic recrystallization.

Recently M. G. Rutten¹ has identified as ignimbrites certain low dome-shaped volcanic cones north of Rome; and has suggested that the fluidization process was involved in their formation. If this claim is substantiated, it provides yet a third way in which the distinctive features of these problem rocks were derived. Among the many occurrences of pyroclasts of this general type the **sillar deposits** which are particularly widely distributed in Peru are of special significance.² They consist of the same material as welded tuffs of other areas—fragments of rhyolite, rhyolitic pumice and various grades of glass—"sand," together with mineral fragments—biotite and rarely quartz, but are largely unwelded. Some are of special interest as *secondary* spherulitic and axiolitic structures have developed after accumulation of the tuff-lava material.

On account of their mode of origin, their occurrence close to active volcanoes, and their fragmental, often finely divided condition, pyroclasts are particularly vulnerable to alteration by circulating volatiles and solutions of volcanic origin. Minerals characteristic of late-stage (deuteric) and hydrothermal processes develop, including low-temperature quartz, chalcedony and opal, also chlorite and, in special circumstances, garnet and late sphene. Of more than general interest are certain agglomerates in the Edinburgh district, Scotland, which contain silicified plant remains. These were thoroughly soaked in colloidal silica so soon after burial that the details of botanical structure are perfectly preserved. Silicification of pyroclastic material in an extremely finely divided condition has resulted in the formation of halleflintas—exceedingly tough, compact and durable rocks, well represented in this country among the Pre-Cambrian volcanics of the Charnwood Forest area.

Little reference has been made in this account to pyroclasts consisting of basaltic glass, for reasons which should be self-evident. Palagonite-tuff falls in this category. The basaltic glass fragments of which it was originally composed have suffered devitrification and conversion into a dull dark green substance of doubtful composition named palagonite (Penck, 1879).

It would be pointless at this stage to attempt to summarize the distribution of pyroclasts in Britain: but they will be referred to as opportunity offers in the last part of this book.

¹ "Ignimbrites or fluidized tuff flows on some mid-Italian volcanoes," *Geol. en Mijnbouw*, New Series, 12 (1959), p. 396.

² Jenks, W. F., and Goldich, S. S., "Rhyolitic Tuff-Flows in Southern Peru," *Journ. Geol.*, 64 (1956), pp. 156-72.

PART IV

**IGNEOUS ACTIVITY IN THE
BRITISH ISLES**

PRE-CAMBRIAN TO DEVONIAN IGNEOUS ACTIVITY

ALTHOUGH it is neither advisable, nor even possible, to divorce stratigraphy entirely from an account of the igneous history of Britain, the details of the former may be largely suppressed except in so far as they bear directly on the sequence of events and the quality of the magma intruded or extruded, as the case may be, during any igneous episode. In this brief account emphasis will be laid only upon those occurrences which are of special interest, in that they have some bearing upon the problems of petrology and petrogenesis.

The igneous cycle when complete comprises:

- (i) a volcanic phase (outpouring of lava),
- (ii) a phase of plutonic (or major) intrusion,
- (iii) a phase of minor intrusion ("dyke-phase").

The products of every cycle are not necessarily completely exposed at the surface at the present time: the active period of the volcanoes may have been of short duration; the amount of lava actually extruded may have been small; subsequent erosion may have been profound. These circumstances have combined in some cases to destroy all record of the volcanic phase. In like manner the plutonic intrusions may have failed to penetrate sufficiently near to the surface to be exposed by subsequent denudation, as happened with the Carboniferous igneous cycle of the Midland Valley of Scotland, the cycle in such cases consisting of lava-flows and minor intrusions only. On the other hand, some cycles were brought to a close by a recrudescence of extrusion.

The most constant phase is that of the minor intrusions. There are, indeed, seemingly unattached dyke-phases that can only with difficulty be assigned to a particular cycle; while individual sills and dykes extend to such great distances from their volcanic centre that there is considerable difficulty in correlating them.

In Britain, volcanic activity played an important role in the Pre-Cambrian, Ordovician, Devonian, Carboniferous and early Permian periods, and in the Tertiary epoch.

PRE-CAMBRIAN

The later stages of the period during which the earth came into being must have been characterized by widespread and intense volcanic activity. The records of this period are lost, but during the Pre-Cambrian epoch lava was injected and extruded on a scale which has seldom been equalled since (e.g. in the Keewatin and Keweenaw in the Canadian Shield).

In the area now occupied by the British Isles, volcanic action, although on a more modest scale than that characterizing the Canadian Shield, must have been widely distributed, since there are few of the comparatively small and isolated outcrops of Pre-Cambrian rocks that are not in part of igneous origin. Outstanding features are the relative abundance of acid lavas (rhyolites); the development at various levels of typical pillow-lavas; the almost invariable presence of sodic plagioclase in place of the calcic varieties more usual in the basic lavas; and the variety and complexity, both in form and composition, of the plutonic masses.

Although, on account of the absence of the usual criteria, correlation of the Pre-Cambrian formations is difficult, it is possible to distinguish an earlier (Archaean) from a later (Proterozoic) period of eruptivity. The former includes the igneous rocks of Anglesey and the coast of Argyllshire, the latter those of Shropshire, the Malvern Hills, Carnarvonshire, St. Davids (Pembroke-shire), the Charnwood Forest (Leicestershire) and some of those in the Channel Islands.

The Lavas

In Anglesey volcanic rocks occur at three chief horizons in the bedded succession of the Mona Complex: in the Fydyln, Amlwch and Gwna Groups basic material predominates, though rhyolitic ejectamenta also occur, chiefly in the lowest group. The basic rocks are typical pillow-lavas, consisting of rounded masses varying from 6 inches to a yard in diameter, and wrapped round by sediment which is sometimes jasper and sometimes limestone. The lavas are obviously submarine, and are accompanied by thin tuff bands. Types that were originally glassy and others that are variolitic spilites are well represented, and have been fully described by E. Greenly.¹ The feldspar is pure albite, while the coloured mineral, common augite, is somewhat better preserved than in most rocks of this kind. Occasional pseudomorphs after olivine also occur. A point of special interest is the discovery that, when traced into regions of intense metamorphism, the normal spilite passes by

¹ "Geology of Anglesey," vol. i, *Mem. Geol. Surv.* (1919), pp. 54 and 71.

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degrees into a rare type of schist consisting essentially of glaucophane and epidote.¹ These glaucophane-schists are restricted to one small area in Anglesey. An outlier of the Mona Complex occurs in the western extremity of the Llyn Peninsula in Carnarvonshire,² and here also spilites are well developed. Plutonic rocks of Pre-Cambrian age form part of the Mona Complex and are referred to below.

In the Tayvallich Peninsula, forming part of the low-lying coastal region of Argyllshire, lavas and contemporary intrusions occur in the Loch Awe Group in the Dalradian.³ That the lavas were submarine is proved by their close association with deep-water sediments, and by the perfect pillow-structure which some of them possess. The multitudinous vesicles are arranged in concentric zones, while pipe amygdales up to one foot in length spring from the base of the flows. These characteristic structures have enabled the succession to be correctly interpreted in spite of the extreme complexity of the tectonics of the region. The lavas were described by Flett as spilites, and consist of albite, chlorite, carbonates and quartz. Closely connected with them, and undoubtedly contemporaneous, are transgressive sills, originally ophitic dolerites, but converted by juvenile reactions into epidiorites consisting of albite, urallite, epidote and spheue.

In Shropshire lavas and tuffs, penetrated by intrusions of leucogranite, form the Uriconian rock groups, occurring both east and west of the plateau of the Longmynd. The eastern belt includes the hills of the Wrekin and Caer Caradoc; while the western culminates in Pontesford Hill, long famous, together with the Lea Rock near Wellington, for their devitrified perlitic and spherulitic rhyolites. Andesitic tuffs and flows of mugearite also occur.⁴ Although the centres of eruption have not been exactly located there is no doubt that these lavas were erupted from volcanoes of the central type.

Lavas probably of nearly the same age constitute the Warren House Volcanic Series in the Malvern Hills. Although poorly exposed they show much the same range of composition as the Uriconian lavas, and include sodic rhyolites, sodic trachytes and variolitic spilites, together with tuffs.

Pebbles of acid lava closely resembling the Uriconian types have been recorded from the Torridonian sedimentary rocks of the North-West Highlands. The northerly provenance of the bulk of the Torridonian sediment, however, renders it highly improbable that

¹ "Geology of Anglesey," vol. i, *Mem. Geol. Surv.* (1919), p. 115.

² Matley, C. A., *Quar. Journ. Geol. Soc.*, **84** (1928), and **88** (1932), p. 238.

³ Bailey, E. B., *Quar. Journ. Geol. Soc.* (1913), p. 280, and the "Geology of Knapdale," *Mem. Geol. Surv.* (1911), p. 59.

⁴ Boulton, W. S., *Quar. Journ. Geol. Soc.*, **60** (1904), p. 470.

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these fragments were carried by coastal drift or otherwise from Shropshire to Scotland. They appear to be the only traces remaining of sheets of rhyolite, the original location of which is unknown. Probably they were contemporaneous with the Uriconian lavas of the more southerly localities.

The **Channel Islands** consist largely of plutonic and schistose rocks with rhyolitic and andesitic lavas locally well developed. The andesites are exposed in the neighbourhood of St. Helier, and on the north-east coast of Jersey. The platy habit and parallel orientation of the feldspars in some types recall the well-known "Carnethy Porphyry" of the Pentland Hills.

The rhyolites are more widely distributed, and are well known on account of their beautiful spherulitic and lithophysal textures.¹ Micro-phenocrysts of quartz and feldspar are sometimes present; flow banding is strongly emphasized by streams of small spherulites, generally the size of a pin's head, but occasionally much larger (Boulay Bay). Some of the thicker flows have a perfect columnar jointing, as near Ann Port. Several of the rhyolites, *e.g.* near Mount Orgeuil, formerly regarded as flows, prove to be sills and dykes penetrating the granite. These rhyolites bear a superficial resemblance to the Uriconian lavas in Shropshire. Similar lavas are well exposed in a belt of country in North Wales lying immediately north of the "slate belt" and crossed by the Llanberis Pass. This is the Padarn Ridge where the rhyolites are seen to be overlain unconformably by the basal Cambrian conglomerates which largely consist of boulders and pebbles of these lavas: they may be late Pre-Cambrian in age.

The Intrusions

Plutonic rocks of Pre-Cambrian age are widely distributed. They comprise a considerable portion of the "Fundamental Complex" of the Archaeozoic of the North-West Highlands,² of the Mona Complex in Anglesey and Llyn, and of the core of the Malvern range. The Lewisian of the North-West Highlands is a Metamorphic Complex originally in part sedimentary, in part igneous. The latter rocks are largely foliated orthogneisses of acid, intermediate, basic and ultrabasic composition. At some period subsequent to their intrusion, but prior to the deposition of the Torridonian, the region was subject to tension acting from north-east to south-west, and a swarm of north-west dykes was injected.³ In these all stages may be traced of the conversion of basalt into

¹ Mourant, A. E., *Quar. Journ. Geol. Soc.*, **89** (1933), p. 273.

² "North-West Highlands," *Mem. Geol. Surv.* (1907), chaps. iii and iv.

³ Sutton, J., and Watson, J., *Quar. Journ. Geol. Soc.*, **106** (1951), pp. 241-308

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hornblende-schist. Typical eclogites occur at Glenelg and show various stages of retrogressive metamorphism into amphibolites.

The Lizard Complex in Cornwall is presumably Pre-Cambrian as are the granodiorites, markfieldites, microdiorites and dacites of the Charnwood Forest in Leicestershire. Of more than usual interest are the so-called **Older Granites of Scotland**. Instead of the compact, laccolithic or batholithic form commonly assumed by granitic intrusions, these bodies are of extreme complexity. They afford an excellent illustration of the process which Harker regards as of the utmost importance in the primary differentiation of igneous rocks, viz. the forcible separation of the liquid from the solid phase during the process of consolidation of a body of magma.¹

The characteristic feature of the more acid types of the older granites is their tendency to subdivision into an infinite number of small intrusions, which often permeate extensive areas and are largely responsible for the high degree of crystallization of the schists and gneisses into which they have been injected.

As a whole they are later than the main Pre-Torridonian folding movements, although the greater part of the intrusion took place when the region was still under the influence of great earth-stresses, with the result that the material of the earlier has often been separated from that of the later consolidation. The latter appears either in the form of minute sills or threads of granite intruded along the divisional planes of the metamorphosed sediments, or as pegmatite in sills or veins. The granite and pegmatite sills are as a rule foliated; while the coarser vein material is more often unfoliated. Probably the intrusions took the form of sills while the district was still affected by the great earth stresses, and that of veins as these stresses gradually ceased.

Even single small intrusions show the result of magmatic differentiation. The central core is often a slightly foliated granite; on the margin a rock of a more alkaline and foliated type predominates; while the fringing veins pass into aplite or pegmatite, which at the tips of the veins passes into almost pure quartz rock.² The complex has evidently resulted from the progressive squeezing out of the liquid portion of granitic magma consolidating under stress. Each portion is slightly abnormal in composition, but were all parts mingled the whole would make up a granite of normal composition. The intrusions can be separated into two great groups by means of the dominant feldspar; in the first (the alkali-granite type) alkali feldspar (orthoclase and microcline) predominates; in the second (the granodiorite type) the alkali feldspar is greatly subordinate to

¹ "Presidential Address," *Quar. Journ. Geol. Soc.*, 73 (1917), pp. lxxvii-xcvi.

² Barrow, G., "Geology of Lower Deeside," *Proc. Geol. Assoc.*, 23 (1912), p. 279.

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oligoclase. Certain of the latter rocks are regarded by H. H. Read as having resulted from interaction between an oligoclase-rich trondhjemitic magma and argillaceous rocks.

In addition to the "Older Granites," however, belonging to the same general period of intrusion (that is, older than the metamorphism of the Highland Schists), basic plutonic rocks in variety also occur. Originally intruded as thick sheets of gabbro and dolerite,

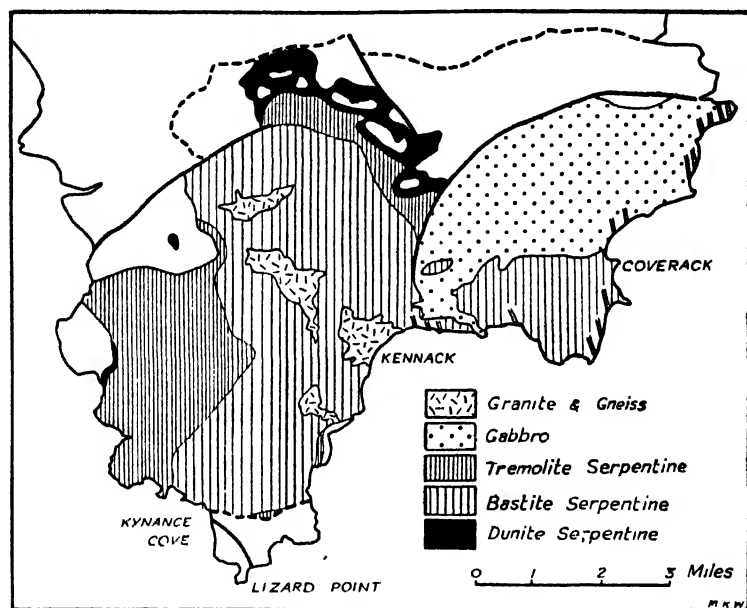


FIG. 130

Sketch-map of the Lizard Complex, Cornwall, after Sir John Flett.
Black dykes indicated around Coverack and Kennack.

they are now largely represented by amphibolites, "epidiorites"¹ and hornblende-schists. One great sill of this kind extends for 50 miles southwards from the coast at Portsoy. This particular sill is noteworthy as it has retained its original characters at Portsoy, where it is associated with anorthosite (an almost unique occurrence for Britain) and serpentinites, representing ultrabasic rocks, the two being in the nature of complementary differentiates.

A plutonic complex presenting many points of interest occurs in the **Lizard** district of South Cornwall (Fig. 130). The rocks into which the complex was intruded are highly metamorphosed sedimentary and hornblende-schists, which were originally tuffs and

¹ Wiseman, J. D. H., "The Central and S.W. Highlands Epidiorites," *Quar. Journ. Geol. Soc.*, 90 (1934), p. 354.

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basic lavas or intrusions, and possibly represent the extrusive phase of the cycle to which the plutonic rocks belong. Evidence of the same extrusive phase is afforded by the hornblende-schists (originally basic lavas) occurring in the Channel Islands and on the mainland of Brittany. Their occurrence in these three isolated localities suggests the possibility of very widespread volcanic activity in the area now occupied by the English Channel.¹

The plutonic rocks of the Lizard² range in composition from peridotites (the earliest intrusions) to alkali-granites (the latest intrusions), the former being dominant. The ultrabasic rocks comprise dunite-serpentinite, tremolite-serpentinite and bastite-serpentinite (which probably originated as bronzite-peridotite). The first two form discontinuous zones round the central core of coarse-grained bastite-serpentinite (Fig. 130). The serpentinite probably has the form of a stratified laccolith, injected during a period of great stress which produced foliation and complicated folded junctions. The serpentinite is penetrated at Coverack by a small intrusion of troctolite, which in turn is cut by veins of gabbro. These are offshoots from a massive intrusion of gabbro which broke through the south-eastern part of the laccolith, and penetrated the serpentinite in a maze of veins and dykes. The coarse-grain of quite thin veins of gabbro proves this later injection to have taken place while the serpentinite was still hot.

After the cooling of the gabbro, a large number of basic dykes (the so-called "black dykes") were injected. Many of these are normal olivine-dolerites, surprisingly fresh for their age, but all stages occur in the conversion of these into hornblende-schists. Others contain significant streaks and blotches of red granitic material which increases in amount in the Kennack gneisses, these having been forced into position under great pressure, and consisting of bands and lenticles of alternating granitic and gabbroic composition.

Finally, red granite-gneiss, similar to the red streaks in the Kennack gneisses, was intruded into the central serpentinite as dykes and small bosses.

The Lizard Complex is the largest mass of serpentinite in the country (20 square miles), and the rocks well illustrate the effects of injection and crystallization under pressure.

The close association of the basic and acid material in the later intrusions points to an equally close relationship when in the molten condition. It is the first of several cases to be mentioned which suggest the co-existence of basaltic and rhyolitic magmas.

¹ Wooldridge, S. W., "Petrology of Sark," *Geol. Mag.*, **62** (1925), p. 241.

² Flett, J. S., "Geology of the Lizard," *Proc. Geol. Assoc.*, **24** (1913), p. 118; and "Geology of the Lizard and Meneage," *Mem. Geol. Surv.*, 2nd Edn., 1946.

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From the **Mona Complex** Greenly has described a series of plutonic rocks ranging from alkali-granites to dunite-serpentinites, and including pyroxenite, gabbro, dolerite and subordinate masses of diorite. The latter are regarded as marginal modifications of the granites.¹ The Sarn granite in Lley, for long regarded as an Ordovician intrusion, is also Pre-Cambrian, and formed part of the ancient land surface on which the basal Ordovician sediments were deposited.²

In **Leicestershire** the Charnian succession³ consists largely of volcanic material, particularly in the lower part of the Middle Group, the Maplewell Series. This includes two spectacular agglomerates: a lower, "Felsitic Agglomerate" and an upper, "Slaty Agglomerate." The latter includes blocks up to five feet in maximum diameter, which indicate proximity to nearby violently explosive volcanoes. These Charnian rocks are associated with a varied suite of intrusions, some few of which are represented by fragments in the agglomerate mentioned above, and their age is therefore definitely fixed. These intrusions include dacite and porphyritic microdiorite (Peldar Tor). The most extensive intrusions in the area are essentially sub-acid, and include the well-known Markfield and Groby rocks. The former gives its name to "markfieldite," a quartz-diorite with micrographic groundmass. The Pre-Cambrian age of these intrusions was proved by quarrying operations in the Nuneaton district, where the basal Cambrian conglomerate is seen to rest unconformably upon an irregular surface of markfieldite, and includes large rounded boulders of the latter.⁴

The third group of intrusives in Leicestershire comprises the much-quarried Mountsorrel granite and its associated more basic derivatives. General agreement as to the age of the complex has not been reached, but it is possibly Caledonian.

ORDOVICIAN

The Cambrian Period was one of quiescence so far as volcanic activity is concerned. With the opening of the succeeding period, however, occurred the first great outburst that ushered in the widespread and long-continued activity of the Ordovician. Some of these early volcanoes were grouped about the northern and southern shores of the geosyncline, which at this time occupied the British

¹ Greenly, E., "Geology of Anglesey," *Mem. Geol. Surv.* (1919), pp. 90-109.

² Matley, C. A., and Smith, B., "Age of the Sarn Granite," *Quar. Journ. Geol. Soc.* (1936), p. 88.

³ Bennett, F. W., Lowe, E. E., Gregory, H. H., and Jones, F., "Geology of Charnwood Forest," *Proc. Geol. Assoc.*, 39 (1928), p. 241.

⁴ Wills, L. J., and Shotton, F. W., "The Junction of the Pre-Cambrian and Cambrian at Nuneaton," *Geol. Mag.* (1934), p. 13.

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area, but others, in the centre of this tract, were localized by movements of elevation—the prelude to the Caledonian revolution.

The eruptions that gave rise to the pillow-lavas (spilites) of Mullion Island off the Cornish coast were possibly contemporaneous with the spilites and dolerites in the Highland Border Series occupying an analogous position on the north of the geosyncline; but the exact age of both series is still a little doubtful.

In early Arenig times eruptions in Pembrokeshire, Carmarthenshire and southern Merionethshire gave rise to the Trefgarn, Llangynog and Rhobell Fawr Volcanic Groups respectively.

While the Trefgarn group consists essentially of andesites, the calc-alkaline lavas of Rhobell Fawr include dacites and basalts, in addition to dominant hornblende- and augite-andesites. This locality is famous for its crystal tuffs, rich in large, though usually broken augites and hornblendes. Rhobell Fawr itself is the basal wreck of the actual volcano, which was of Vesuvian type, being built of agglomerates, tuffs, lava-flows and vent-intrusions. Emanations connected with the waning phase of activity effected a complete pseudomorphism of most of the original minerals, and have riddled the rocks with epidote and pyrite. Numerous sills crowded together west of (originally beneath) the volcanic pile indicate a well-developed "dyke-phase" to which the lithologically similar sills, dykes and small laccoliths occurring in the Harlech Dome and intimately associated with the gold-bearing quartz veins in the Dolgelley district,¹ undoubtedly belong. The rocks of the minor intrusions are those that should be associated with andesitic lavas, being hornblende-microdiorites of porphyritic and aphyric types.²

In later Arenig times volcanic action became more general. To this episode belong the Coomb Volcanic Series of Llangynog in Carmarthenshire,³ the Skomer Island Volcanic Series, forming the island of that name off the coast of Pembrokeshire,⁴ and the Lower Acid Series of Merionethshire. At the same time in the zone of *D. extensus*, submarine eruptions in the Southern Uplands of Scotland gave rise to the Ballantrae and Sanquhar Volcanic Series in the Girvan district.⁵ Of these several series, those in Merioneth and Carmarthen consist essentially of acid lavas and tuffs, those of Girvan are dominantly albite-basalts (spilites), with well-developed pillow-structure, magnificently exposed on the coast near Ballantrae and associated with coarse agglomerates. In the Skomer Island

¹ Dolgelley is now officially spelled Dolgellau.

² Wells, A. K., *Quar. Journ. Geol. Soc.* (1925), p. 463.

³ Cantrill, T. C., and Thomas, H. H., *Quar. Journ. Geol. Soc.*, **62** (1906), p. 223.

⁴ Thomas, H. H., *Quar. Journ. Geol. Soc.*, **67** (1911), p. 175.

⁵ Balsillie, D., "The Ballantrae Complex," *Geol. Mag.* (1937), p. 177, but cf. *ibid.*, p. 188.

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Series, soda-rhyolites, soda-trachytes (keratophyres), trachybasalts, mugearites and normal olivine-basalts occur as flows.

Following a short period of quiescence during which normal sedimentation took place, the Llanvirn period witnessed the maximum of Ordovician volcanic activity; and simultaneous eruptions occurred at several centres in North and South Wales,¹ South Shropshire and the English Lake District. In each of these localities the Llanvirn volcanic rocks rest upon blue-black shales containing the *Didymograptus bifidus* fauna, which accurately dates the eruptions and proves them to have been submarine. It is possible, however, that local movements of uplift and the accumulation of great thicknesses of lava and ash enabled some of these volcanoes to emerge above sea-level and become volcanic islands. In South Wales the volcanoes soon became extinct; in Shropshire activity ceased in *bifidus* times but was renewed in the Caradocian period; in North Wales they were continuously active in one locality or another until late Caradocian times. Towards the close of the period represented by the zone of *Nemagraptus gracilis*, the volcanoes of southern Merionethshire became finally extinct, but new eruptive centres were established northwards in Snowdonia and southwards in the Wells country of East-Central Wales (Llanwrtyd Wells,² Llandrindod and Builth).³ The lavas and tuffs of Carnarvonshire were erupted from a series of aligned cones extending south-westward from the north coast at Conway. They have long been known as the "Bala Volcanic Series." The Llanvirn and later igneous activity will be considered under the several districts concerned.

North Wales. (a) The Volcanic Rocks.—The Ordovician volcanics in North Wales are of three different kinds: rhyolites (potassic and sodic), andesites and spilites. A noteworthy feature is the great development of pyroclasts of all degrees of coarseness. These occur on all the volcanic horizons, and usually bulk far larger than the actual flows. The basic lavas are spilites, characterized by pillow structure which attains its most perfect development in the Lower Basic Group near Arthog (Merionethshire) and in the Upper Basic Group on Cader Idris. The former has a maximum thickness of 1,500 feet and the latter of about 500 feet. In addition to typical spilitic pillow lavas these Basic Groups include bomb-tuffs, vitric tuffs, feldspar-crystal tuffs and rocks in which pyroclastic and sedimentary materials are mixed in all proportions.

East of Rhobell Fawr a variolitic basalt of normal composition

¹ Thomas, G. E., and Thomas, T. M., "Volcanic Rocks of the Area between Strumble Head and Fishguard," *Quar. Journ. Geol. Soc.*, **112** (1956), pp. 291-314.

² Stamp, L. D., and Wooldridge, S. W., *Quar. Journ. Geol. Soc.*, **79** (1923), p. 16.

³ Jones, O. T., and Pugh, W. J., *Quar. Journ. Geol. Soc.*, **104** (1948), p. 71.

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(labradorite, augite and ilmenite, with the usual decomposition products) occurs in close association with typical spilites. When followed to the north, the spilites give place to hypersthene-andesites, which were erupted from the Arenig Mountain centre.¹ The fact that the local volcano had built itself up above sea-level accounts for the absence of pillow structure; but in addition the lavas are of andesitic, not spilitic, type.

It is significant that spilitic pillow-lavas associated with a rhyolitic lava and pyroclasts in the *hirundo* and *bifidus* zones occur in South-West Llyn in the Carnarvonshire peninsula;² but in Anglesey and Snowdonia no volcanic rocks occur on these horizons.

In the Cader Idris district acid lavas and tuffs were erupted in post-*hirundo*-pre-*bifidus* times, and again towards the close of the volcanic episode, forming in the latter case the Upper Acid Group, with a maximum thickness of 1,500 feet,³ which includes lavas of variable composition: some are potassic rhyolites, others sodic (quartz-keratophyres) while some are less rich in SiO₂ and are trachytes and keratophyres (albite-trachytes). To the north this group persists into the Arenig Mountain district.

The great majority of the rhyolitic lavas of Carnarvonshire are of Bala age. They were grouped by Harker⁴ into four series, to which he gave local names. Much of the area has been resurveyed by Howel and David Williams,⁵ who have shown that no lavas occur on Snowdon below the zone of *Nemagraptus gracilis* (the zone to which the highest lavas of Cader Idris and Arenig belong). Further, a thickness of some 1,600 feet of unfossiliferous, and therefore undated, strata intervene between the fossiliferous horizon and the actual base of the Snowdonian volcanics. In the Dolwyddelan Syncline east of Snowdon the highest volcanics are immediately overlain by black graptolitic slates, in the zone of *D. clingani*. It follows that the volcanic episode was of relatively short duration—perhaps within the period of one zone.

The general succession on Snowdon comprises the three divisions:

3. Upper Rhyolitic Series.
2. Bedded Pyroclastic Series.
1. Lower Rhyolitic Series.

¹ Fearnside, W. G., *Quar. Journ. Geol. Soc.* (1905), pp. 608–40.

² Matley, C. A., "Geology of Mynydd rhiw," *Quar. Journ. Geol. Soc.*, **88** (1932), p. 238.

³ Davies, R. G., "The Cader Idris granophyre and its associated rocks," *Quar. Journ. Geol. Soc.*, **115** (1959), p. 189.

⁴ "The Bala Volcanic Series of Carnarvonshire," *Sedgwick Prize Essay for 1888* (1889), p. 9.

⁵ Williams, H., "Geology of Snowdon," *Quar. Journ. Geol. Soc.*, **83** (1927), p. 346; Williams, D., "The Geology . . . of Nant Peris . . .," *Quar. Journ. Geol. Soc.*, **86** (1930), p. 191; Williams, H., and Bulman, O. M., "Geology of the Dolwyddelan Syncline," *Quar. Journ. Geol. Soc.*, **87** (1931), p. 425.

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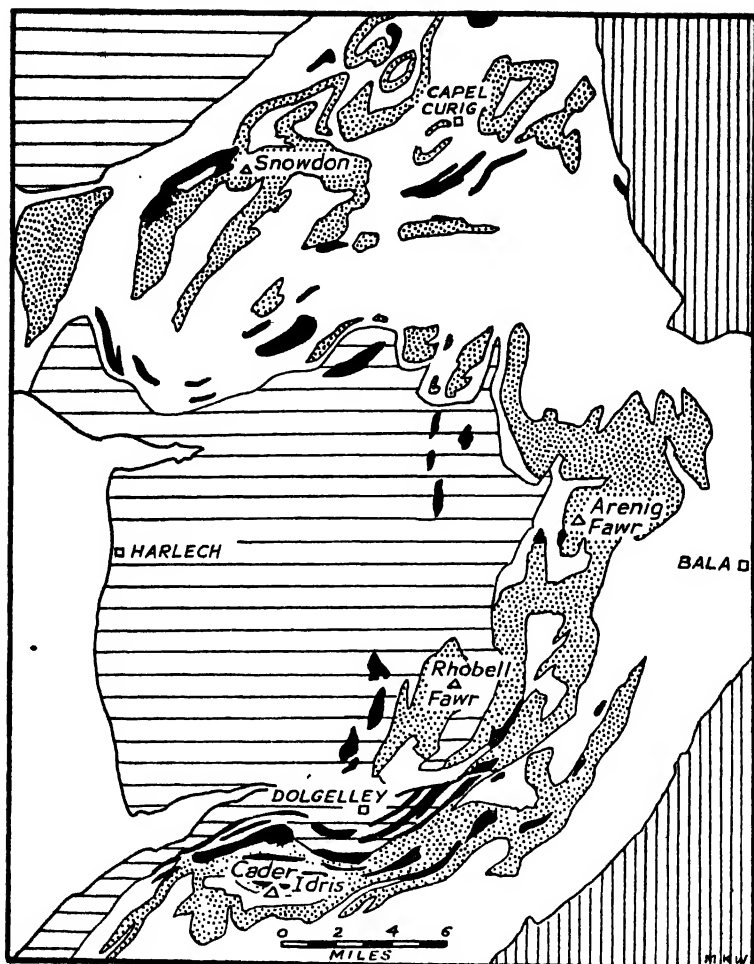


FIG. 131

Sketch-map of part of North Wales showing the distribution of the igneous rocks: volcanic, stippled; intrusive, black; Cambrian, horizontal ruling; Ordovician, plain; Silurian, vertical ruling. After *Geol. Surv.*, A. H. Cox, A. K. Wells and others.

The Lower Rhyolitic Series commences with a nodular rhyolite, succeeded by well-banded rhyolites in which more quartzose and more feldspathic bands alternate. Where subjected to pressure they pass rapidly into quartz-sericite-schists. The only coloured minerals are green and brown biotite, and the lavas are usually rich in the garnet, spessartine. Locally more than three-fourths (1,500 feet

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maximum) of the Lower Rhyolitic Series consists of pyroclastic rocks, first lithic, then crystal, and finally vitric tuffs, all of the composition of potassic rhyolite. The pyroclasts are essentially glass-dust rocks, now termed ignimbrites.

The Bedded Pyroclastic Series includes a great variety of lavas and pyroclasts reaching 1,200 feet in thickness. In the main they are basic pumice-tuffs interbedded with spilites and rare flows of rhyolite. Of special interest is the occurrence of numerous bombs of spilite and keratophyre especially abundant about Snowdon's summit. The presence of acid lavas associated with these basic rocks indicates that more than one group of vents was in operation at this period, which was that of maximum differentiation of the magma.

The Upper Rhyolitic Series consists of potassic rhyolites, though rather richer in soda than the earlier acid flows. They are poor in phenocrysts, which include orthoclase and acid plagioclase, while biotite is commonly present.

(b) **The Intrusive Rocks.**—The Lower Palaeozoic rocks of Wales are penetrated by large numbers of intrusions. On the grounds of petrographical similarity it is clear that the hornblende-microdiorites of Merionethshire represent the hypabyssal phase of the Rhobell-eruptions. Similarly the hypersthene-andesites intrusive into the rocks building Arenig Mountain must have been nearly contemporaneous with the eruptions from that centre. In both cases there is often difficulty in determining whether a particular rock is extrusive or intrusive. But in addition to these local types, there are dolerite sills of regional distribution and uncertain date. Although there were no Carboniferous or Tertiary volcanoes in Wales, intrusions of both these ages occur in Anglesey, Carnarvonshire and possibly farther south. Intense folding also took place in Wales during the Caledonian revolution, and it is possible that these earth-movements were accompanied by intrusive phenomena. Nevertheless, most of the intrusive rocks, other than those of Pre-Cambrian age, must be regarded as probably belonging to the Ordovician cycle. The influence of the local centres is shown even in the regional sills: those in the Arenig district show affinity with the lavas by containing hypersthene, while the feldspar is stated to be less basic than in the dolerites—andesine in place of labradorite. In those areas where the facies of the extrusive phase was spilitic, the dolerites are usually albitized, while the spilitic dolerites (which were intruded into soft mud on the sea-floor and reproduce many of the characters of the lavas) are the connecting link between the two phases. The normal dolerites are of very simple mineral composition, being plagioclase, augite, ilmenite rocks, with ophitic and ophimottled textures. The central parts of the more massive intrusions are coarse-grained and gabbroic. None of these rocks contains olivine. Quartz-dolerites and

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spilitic dolerites both occur in Snowdonia and in South Merionethshire. Although in eastern Carnarvonshire the dolerites are often fresh, it is rare to find any original minerals among those of Snowdonia, where they are albitized and chloritized. In Merionethshire the dolerite intrusions are especially developed in the neighbourhood of Dolgelley, occurring chiefly as sills, particularly in the Basement Group of the Ordovician. In Snowdonia, the doleritic intrusions are dykes in the Cambrian rocks, but sills and small phacoliths in the Ordovician volcanics.

Numerous concordant intrusions of albite-dolerite occur in the Builth-Llandrindod Ordovician inlier in Mid-Wales.¹ They seem to demonstrate the influence of depth of intrusion upon the form of the rock-bodies, which vary from sills in the Llanvirnian to small, widely scattered bun-shaped masses in the Llandeilian. Some appear to be laccoliths, with feeders, of the conventional type.

Large numbers of intrusions (often composite) occur in the south-west portion of the Llyn Peninsula, and are intrusive into the *D. bifidus* and lower beds. Petrographically these are linked with the Palaeozoic sills, dykes and laccoliths of Anglesey. Distinctive features are the occurrence of hornblende in even the most basic types, and the large amount of olivine sometimes present. The rock-types represented include minverite (hornblende-dolerite), olivine-dolerite and hornblende-picrite. In mineral composition these rocks appear to be distinct from those occurring farther south, and Greenly claims them to be of late-Caledonian age.

The acid intrusions of Snowdonia followed the eruptive phase, and include representatives of all the commoner acid minor intrusive rocks. Phenocrysts of quartz, micropertthite and orthoclase are frequently plentiful, and are embedded in a micro- to cryptocrystalline groundmass of soda-orthoclase and quartz. These acid minor intrusions are petrographically allied to the rhyolite flows. The granophyres (graphic microgranites) are characterized by the presence of augite, while the microgranites contain both this mineral and biotite. In two localities, Mynydd Mawr and Bwlch-y-Cywion, the place of these minerals is taken by riebeckite (Fig. 84). Mynydd Mawr is a boss-like mass three miles west of Snowdon. It is probably the plug of a volcanic vent of Bala age, marking the source of some of the lavas of Snowdon, Moel Hebog and Llwyd Mawr. A similar riebeckite-microgranite with a flow-banded rhyolitic marginal facies occurs near Llanbedrog, in the Llyn Peninsula.² At Tan-y-Grisiau, 2 miles north of Ffestiniog, there is an intrusive

¹ Jones, O. T., and Pugh, W. J., "A Multi-layered Dolerite Complex . . .," *Quar. Journ. Geol. Soc.*, 104 (1948), p. 43; and later paper, *ibid.*, p. 71.

² Matley, C. A., *Quar. Journ. Geol. Soc.*, 94 (1938), pp. 596 and 603.

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mass of microgranite (in places micrographic), $2\frac{1}{2}$ miles long. The date of the intrusion was subsequent to the hardening and cleavage of the Tremadoc rocks and may be much younger than the Ordovician. The rock is characterized by abundant quartz and poorness in ferromagnesian minerals. The feldspar appears to be largely a triclinic variety, possibly soda-orthoclase.¹

Acid intrusive rocks (granophyres) in southern Merionethshire form two massive columnar sills in the neighbourhood of Dolgelley, one building the precipitous scarp of Cader Idris, and the other occurring in the foothills to the north.² Both very closely resemble certain of the volcanic rocks nearby, and doubtless represent rhyolitic magma which failed to reach the surface and consolidated under a sedimentary cover.

Few of the larger intrusions in Carnarvonshire have been studied in detail: those that have indicate their composite nature, due in some cases to successive injections and in others to differentiation in place. An example of the latter is Penmaenmawr Mountain between Conway and Bangor.³ The intrusion becomes progressively more basic from the summit downwards. The upper part is a graphic microdiorite approaching markfieldite, while the rock building the lowest visible portion is nearly doleritic in composition, being a labradorite-pyroxene rock, with the former mineral in excess.

Shropshire and the Welsh Borders.—Ordovician andesites occur in the hills of the Shelve district, near the boundary between Shropshire and Montgomeryshire⁴ (Fig. 132). Here lavas of lower Llanvirn age form the Stapeley Hills, attaining their greatest thickness and extent on Todleth. These lavas, tuffs and associated intrusive andesites contain phenocrysts of plagioclase, augite, and hypersthene in a fine microlitic groundmass. In the same area thick sheets of similar composition are intercalated in the Bala rocks.

Precisely similar hypersthene-andesites occur in the Breidden Hills (e.g., at Moel-y-Golfa, 6 miles north-east of Welshpool), which constitute an outlying portion of the Shelve area. Vesicular andesites and tuffs also occur near Builth, in the range of the Carneddau (Herefordshire and Radnorshire), where an outlier of Ordovician rock occurs.

The Ordovician rocks of Llanwrtyd Wells include volcanic material of acid and basic composition: the acid rocks are fragmental soda-rhyolites, some showing perlitic texture, while the basic rocks are typical spilites showing pillow-structure. They occur in the zone of

¹ Jennings and Williams, G. J., *Quar. Journ. Geol. Soc.*, 47 (1891), p. 368.

² Cox, A. H., and Wells, A. K., "The Lower Palaeozoic Rocks of the Arthog District (Merionethshire)," *Quar. Journ. Geol. Soc.*, 76 (1921), pp. 283-305. See also Davies, R. G., *op. cit.*, p. 437.

³ Sargent, H. C., *Proc. Geol. Soc. Liverpool*, 14 (1924), p. 82.

⁴ Whittard, W. F., *Proc. Geol. Assoc.*, 42 (1931), p. 322.

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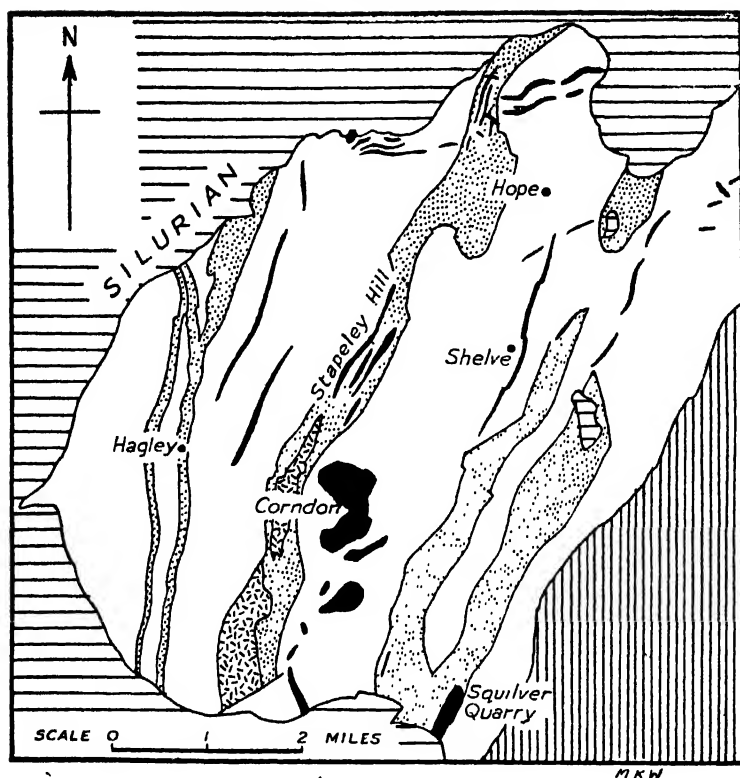


FIG. 132

Sketch-map of the Ordovician inlier of the Shelfe area, Shropshire, showing the outcrops of the lavas and associated intrusions. After W. W. Watts and F. G. Blyth. Volcanic rocks, stippled; basic-ultrabasic, black; intrusive andesites, small dashes; Ordovician sediments, blank.

Nemagraptus gracilis, and are therefore contemporaneous with the Upper Basic and Upper Acid Groups of Cader Idris.¹

The comparative uniformity in the petrographical character of the lavas is reflected in the minor intrusives, of which two main types occur. The first of these are intrusive andesites, found in the area of the Stapeley Volcanic Series on Stapeley Hill, Llanfawr, Roundtain, and Todleth.² In the Breidden Hills they form the greater part of Moel-y-Golfa. From their lithological characters

¹ Stamp, L. D., and Wooldridge, S. W., *Quar. Journ. Geol. Soc.*, **79** (1923), p. 16.

² Watts, W. W., "The Geology of South Shropshire," *Proc. Geol. Assoc.*, **36** (1925), p. 359.

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some were formerly thought to be lavas. The second type is coarse-grained dolerite, resembling the gabbroid dolerites of North Wales. The Ordovician rocks of the Shelve area are pierced by many dykes and sills of dolerite.¹ At the Corndon and at Pitchfolds the phacolithic character of these intrusions is apparent. At the former locality the phacolith is intrusive into the *bifidus*-shales, a little below the overlying Stapeley Volcanic Series. In the Breidden Hills, hypersthene-dolerites containing both hypersthene and augite, and often with ophitic texture, occur at Rodney's Pillar, Criggion, Belau Bank, and Trewern, and are similar in form to the Corndon phacolith. In addition to the basic intrusives small outcrops of augite-picrite occur in the Shelve district near Shelve Church and Cwm-mawr. Some doubt exists (in the absence of direct evidence) as to their age; but as augite-picrite would be the normal, more basic differentiate of a magma which gave rise to the dolerites, and since the two types occur together in the minor intrusions of Carnarvonshire, it seems probable that the Shropshire picrites are of the same age, and belong to the Ordovician cycle.

The dolerites of this area are not all of the same age: cognate xenoliths of dolerite are found in the intrusive andesites referred to above. The latter, on Moel-y-Golfa, were definitely emplaced and exposed by denudation prior to the deposition of the unconformable Llandovery (Silurian) rocks. The dolerites of the Carneddau range, near Builth in South-West Radnorshire, cut Ordovician (Llandeilo) rocks, but not the Llandovery beds. Consequently some, at least, of the dolerites must be of Ordovician age. Blyth has shown that the Squilver gabbro (Fig. 132) formed a "feature" on the early-Silurian coastline, as pebbles of this rock occur in the Upper Valentinian beach-deposits forming the local base of the Silurian. As this gabbro metamorphoses Llanvirnian shales, the age of these intrusives is proved.

The Lake District.²—The vast accumulation of volcanic material known as the Borrowdale Volcanic Series is mainly made up of andesitic lavas and tuffs; but, as in the case of the contemporaneous cycle in North Wales, a change in the character of the magma took place towards the close of the volcanic episode, the andesites and andesitic pyroclasts being succeeded by rhyolites including welded tuffs.³

The great thickness originally assigned to the Borrowdale Volcanic

¹ Blyth, F. G. H., "Intrusive Rocks of the Shelve Area," *Quar. Journ. Geol. Soc.*, **99** (1943), p. 169.

² Green, J. F. N., "Vulcanicity of the Lake District," *Proc. Geol. Assoc.*, **30** (1919), p. 153; also Mitchell, G. H., "Borrowdale Volcanic Series, etc.," *Quar. Journ. Geol. Soc.*, **90** (1934), p. 418.

³ Oliver, R. L., "Welded Tuffs in the Borrowdale Volcanic Series . . .," *Geol. Mag.*, **91** (1954), pp. 473-83.

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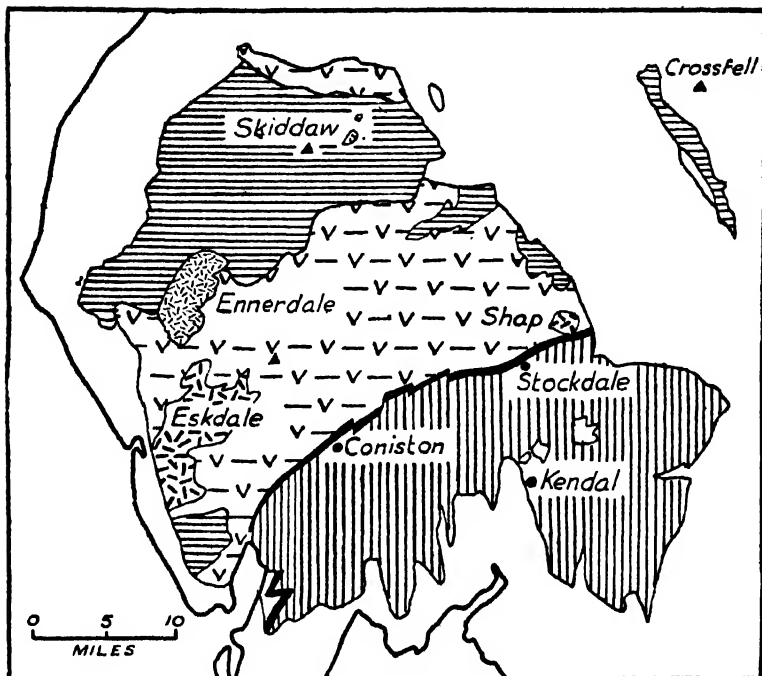


FIG. 133

Sketch-map of the English Lake District, showing the major rock-groups and the more important intrusives. Ordovician, horizontal ruling; Silurian, vertical ruling; Borrowdale Volcanic Group, V.

Series has been proved to be considerably over-estimated; but they probably exceed 10,000 feet. During the Caledonian revolution they were subjected to very complex folding and profound faulting.

The succession commences with true explosion tuffs, formed of material ejected during the first paroxysmal outbursts. Above these, pyroxene-andesites are followed by tuffs, and these in turn by andesites and rhyolites. The lavas are frequently autobrecciated, this being accentuated on weathered surfaces, and causing a deceptive resemblance to coarse agglomerates. Petrographically the lavas are very variable, both in texture and in the proportions of the component minerals, and rock types other than the dominant andesites may be present.

The Eycott lavas that occur at Eycott Hill, one mile from Troutbeck Station, near Keswick, and in the Cross Fell range, constitute a more basic type, probably more correctly classified as basalts. Some of the members of this series are strongly porphyritic, containing

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fine large crystals of plagioclase (bytownite to anorthite), also idiomorphic hypersthene altered to bastite.

Rhyolites form the highest part of the Borrowdale Volcanic Series. They are felsitic rocks, presenting strong flow structure, and occasionally containing garnets (as at Illgill Head). They occur in Langdale, on Crinkle Crag and Great Gable (Sty Head), in Long Sleddale and near Great Yarlside. The basal flow is a typical nodular rhyolite, which has been traced over a considerable area, but several of the rhyolites are intrusive, and welded tuffs occur.

The lowest lavas in the main Borrowdale Series, which appear to succeed the Lower Llanvirn conformably, must have been submarine, and poured out in deep water. Shallowing of the water is indicated by the strongly marked false-bedding at higher levels in certain of the tuffs.¹

In addition to the main outcrops, the Borrowdale volcanics are represented in the Cross Fell inlier by a small thickness of andesitic lava and tuff (the Milburn Group) in the *bifidus* zone,² as well as by the overthrust masses referred to above. In the Sedbergh inlier rhyolitic flows occur in the Ashgillian.

The Lake District has its plutonic rocks in addition to a host of minor intrusives. Several of the intrusions occur at the junction of the Skiddaw Slates with the overlying Borrowdale Volcanic Series. The Eskdale Granite (45 square miles in area), the largest intrusion in the district,³ the Buttermere and Ennerdale granophyre, the St. John's Vale microgranite and the gabbro-granophyre sheet-complex of Carrock Fell all occur at this horizon. Other masses consolidated before penetrating so high: thus the famous Shap Granite occurs near the junction of the Borrowdales and the Conistone Limestone Series.

It forms three isolated valley inliers, but its true size is indicated by the very extensive metamorphic aureole within which the slates have suffered not only thermal metamorphism but also metasomatism. The granite itself has been modified profoundly by late-stage solutions which accumulated near the roof of the intrusion and have converted the northernmost outcrop of granite into the largest mass of greisen in Britain—the Grainsgill Greisen. The picrite of Great Cockup and many dolerite sills also occur in the Skiddaw Slate.

Very different opinions are held as to the age of many of these intrusions: the great granite masses of Shap, Skiddaw and Eskdale

¹ Hartley, J. J., *Proc. Geol. Assoc.*, 36 (1925), p. 203.

² Shotton, F. W., "The Cross Fell Inlier," *Quar. Journ. Geol. Soc.*, 91 (1935), p. 149.

³ Simpson, B., *Proc. Geol. Assoc.*, 45 (1934), p. 17.

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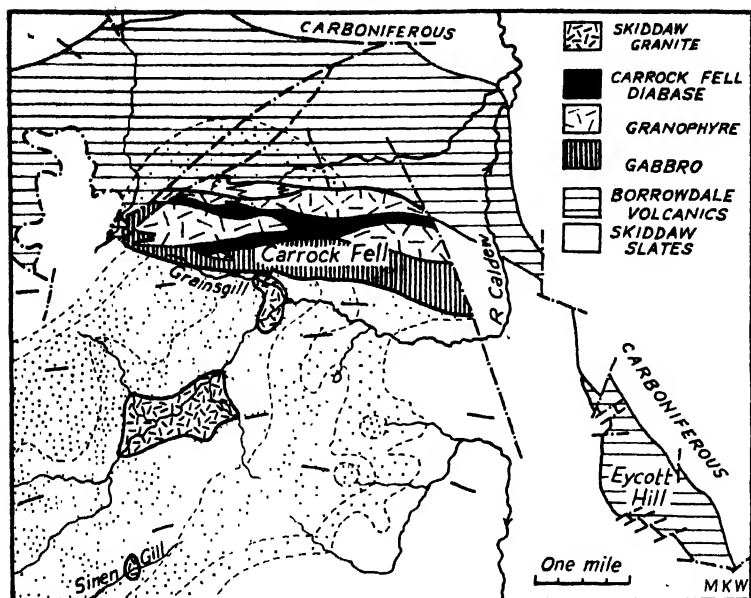


FIG. 134

Carrock Fell Complex and the Skiddaw Granite, Cumberland.
(After S. E. Hollingworth.)

Three zones of the metamorphic aureole of the granite are shown by stippling in the region of Skiddaw Slates. Differences of metamorphic grade cannot be distinguished in the Carrock Fell Complex which is therefore shown without stipple. Fold-axes in the Skiddaw Slates are indicated.

are commonly regarded as being of Caledonian age. This is certainly true of the Shap intrusion, but an attempt has been made to show that the other two may belong to the Ordovician cycle.¹ Similarly, the age assigned to the Carrock Fell Complex varies according to different authorities from pre-Bala to Tertiary. Much depends upon the identification of certain rock-fragments occurring in the basal conglomerates of the Conistone Limestone Series.

Petrographically the Lake District intrusives cover a very wide range of composition—from very acid granophyres and granites to ultrabasic picrites. Acid and basic types again predominate: there are few of intermediate composition. As in North Wales, the granophyres pass, towards their margins, into more basic modifications approaching markfieldite in composition (Ennerdale).²

¹ Green, J. F. N., *Proc. Geol. Assoc.*, **29** (1918), pp. 126–36, and **28**, pp. 1–30.
² Rastall, R. H., *Quar. Journ. Geol. Soc.*, **62** (1906), p. 253.

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In the Carrock Fell Complex¹ (Fig. 134), north-east of Keswick, the most interesting features are the steeply dipping sheeted form of the complex, the close association of gabbro and granophyre; the presence of hybrid zones involving the two main types; and the differentiation of the gabbro. The latter is mostly quartz-bearing and hornblendic, while it is abnormally rich in ilmenite which increases towards the margins until ultimately it forms a considerable proportion of the rock.

Of the many minor intrusions, the following may be mentioned. The Armboth dyke is a spherulitic, or granophyric quartz-porphyry, composed of bright-red feldspar (orthoclase) and dark bipyramidal quartz crystals scattered evenly through a dun-coloured groundmass containing garnets. Porphyritic microgranites occur as bosses and dykes: thus the two laccoliths of St. John's Vale, one of which is quarried near Threlkeld Station, are of this type, containing phenocrysts of orthoclase and small garnets; and dykes of a similar rock, with or without porphyritic crystals of quartz and feldspar, occur, for instance, at a number of localities. In the Westwater district there are, connected with the Eskdale and the Ennerdale masses, innumerable dykes and sills, some of which contain feldspar phenocrysts up to two inches in length. A remarkable porphyritic microgranite containing, besides phenocrysts of a red orthoclase, plagioclase and quartz, large plates of muscovite and small flakes of biotite occurs at Dufton Pike in the Cross Fell inlier, and is known locally as the "Dufton Granite." A dyke of spherulitic felsite traverses the rocks of High Fell in Cumberland.

In the areas considered above the intrusions are closely associated with lavas of essentially the same petrographic type; but in other tracts where Lower Palaeozoic rocks occur this is not the case. For example, in the Cambrian inlier lying west of the Malvern Hills² several sills and small bosses occur. They are not accompanied by lavas, and direct proof of their age is wanting. It is significant, however, that no intrusions occur in the adjacent Silurian strata, so that an Ordovician age is implied. This supposition is strongly supported by the facies of the rocks, which are spilitic. The types represented are spilitic andesites, vesicular spilites and spilitic olivine-dolerites.

The Loch Borolan Complex.³—An interesting intrusive complex

¹ Hollingworth, S. E., "Carrock Fell and adjoining areas," *Proc. Yorks. Geol. Soc.*, 23 (1938), pp. 208-18.

² Blyth, F. G. H., "On the Intrusive Rocks in the Cambrian inlier near Malvern," *Quar. Journ. Geol. Soc.*, 91 (1935), p. 463.

³ Shand, S. J., "Loch Borolan Laccolith, North-West Scotland," *Journ. Geol.*, 47 (1939), p. 408.

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occurring in the Assynt district of Sutherlandshire cuts and metamorphoses Cambrian rocks. It is affected by the Caledonian thrusts, and hence is probably of Ordovician age. In facies it differs from other intrusive rocks of this period, and, indeed, contains rock types unique in this country and uncommon elsewhere.

The complex has the form of a stratified laccolith, the members of which grade one into another, and hence were derived by differentiation in place. The highest rock exposed is a quartz-bearing soda-syenite (nordmarkite). Beneath this, megascopic quartz is absent, and with its disappearance nepheline and its alteration products come in, while the proportion of coloured minerals to perthitic feldspars gradually increases. The central zone of the laccolith is occupied by a melanite-nepheline-syenite, while the lowest portion exposed is a melanite-nepheline-gabbro. It is probable that the base of the intrusion (which is hidden) consists of a garnet-rich pyroxenite.

The minor intrusions connected with the Loch Borolan Complex are not less interesting and include nepheline-syenite-pegmatites, aegirine-aplites, pseudoleucite-porphry and the problematical rocks, borolanites. These rocks were believed to contain pseudomorphs after leucite, since they resemble very closely undoubted pseudoleucites of other (American) localities, but Shand has thrown doubt upon this identification.¹

At two other localities in northern Scotland syenitic complexes occur which, from their petrographic characters, are obviously comagmatic with the Loch Borolan Complex. Indeed one of them, the Loch Ailsh Complex, is only two miles distant from the latter, and like it, is intrusive into the Cambro-Ordovician rocks outcropping to the west of the great overthrust faults which bound the North-West Highlands. The Ben Loyal alkali-complex, on the other hand, builds a picturesque mountain group lying ten miles east of the overthrust zone, in the heart of the Moine Schists of uncertain, but presumably Pre-Cambrian, age.

The Loch Ailsh mass² has the form of a sheeted complex or stratiform laccolith, consisting essentially of sodi-potassic syenites. The highest member contains quartz, and approximates closely to nordmarkite. Downwards, quartz fails, as in the Loch Borolan Complex, and the dominant type is pulaskite, with colour index of 10 or over, the mafic components including aegirine and melanite garnet, with riebeckite in one variety. Still lower, the proportion of

¹ "On Borolanite and its Associates in Assynt," *Trans. Edin. Geol. Soc.*, 9 (1909-10), pp. 202-15 and pp. 376-416. See also: Tilley, C. E., "Some new chemical data on assemblages of the Assynt alkali suite," *Trans. Edin. Geol. Soc.*, 17 (1958), pp. 156-64.

² Phemister, J., in "Geology of Strath Oykell and Lower Loch Shin," *Mem. Geol. Surv.* (1926), p. 22.

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coloured minerals increases, reaching about 70 per cent in a shonkinite zone, which in turn passes down into an ultrabasic layer including biotite-pyroxenite, biotite-hornblende-pyroxenite and hornblendite. This major portion of the complex is cut by a later intrusion of ultra-feldspathic syenite, termed *perthosite* by J. Phemister, on account of its extraordinarily high content of anti-perthitic feldspars.

The **Ben Loyal Complex** has been shown by H. H. Read¹ to consist of rock types closely comparable with the Loch Ailsh syenites. Pulaskite makes up the greater part of the 12 square miles exposed, but nordmarkite is also well represented.²

The three complexes, together with the associated minor intrusions, comprise an alkali-province of unique type so far as the British Isles are concerned.

SILURIAN

Igneous activity in the Silurian period was restricted to feeble outbursts in one Irish (Clogher Head), one Welsh and two English localities. In the **Tortworth Inlier**, lying north of the Bristol Coal-field, two bands of igneous rock occur in Silurian strata. The higher of these is considered to be a lava-flow, since it is accompanied by tuffs, while the lower is probably intrusive. The lava is an enstatite-andesite, containing many corroded xenocrysts of quartz which, as usual, are surrounded by reaction rims of pyroxene. The rocks of the lower band are more basic, and appear to be intrusive albitized olivine-basalts.³

The only other English locality where lavas of Silurian age are known to occur is near Shepton Mallet in the eastern parts of the **Mendip Hills**. The rocks, extensively quarried for road-metal, are purple and green andesites and andesitic tuffs. They contain conspicuous phenocrysts of plagioclase and bastite pseudomorphs after rhombic pyroxene.⁴

Silurian volcanic rocks occur in West Pembrokeshire at Marloes, in Upper Valentian strata. They are olivine-basalt flows showing pillow-structure.

The lavas of **Clogher Head** in County Kerry differ from the other Silurian examples by being acid in composition and by including nodular and banded rhyolites. The associated sediments are of Llandovery and Wenlock age.

¹ Read, H. H., "Geology of Central Sutherland," *Mem. Geol. Surv.* (1931), p. 174.

² King, B. C., "The Cnoc nan Cuilean area of the Ben Loyal Complex," *Quar. Journ. Geol. Soc.*, **98** (1942), p. 147.

³ Reynolds, S. H., *Quar. Journ. Geol. Soc.*, **80** (1924), pp. 106-11.

⁴ *Ibid.*, **63** (1907), p. 218.

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Mention may be made of the "green streak" in the zone of *Monograptus argenteus*, in the Skelgill Beds of the Llandoveryian of north-western England and central Wales. This interesting band is believed to represent a distant volcanic eruption, and consists of the finest wind-blown volcanic dust.

For the rest the Silurian was a period of tranquillity: it was the calm before the storm, however, for with the close of the period came the main phase of the Caledonian revolution, which is considered in relation to contemporary igneous activity in the following pages.

THE CALEDONIAN INTRUSIONS

A small-scale geological map of Scotland shows a large number of major intrusions, some of very large size. These have been termed the "Newer Granites" to distinguish them from the "Older Granites," which, as we have seen, are of Pre-Torridonian age. A better name would be "Newer Intrusives," for many rocks in addition to granites are included. It will be observed that the intrusions are aligned roughly parallel to the great faults which divide the country into regions and have the direction of the Caledonian mountain chains (the Caledonides of Suess). Doubtless the granites worked their way up into the cores of the mountains and have been exposed by subsequent profound denudation. As the intrusions came into existence at the same time as the mountain chains, it is best to refer to them as the Caledonian intrusions. They are not restricted to Scotland and England: others are found in the prolongation of the Caledonian chains in Ireland (the Newry granite, for example)¹ and Scandinavia.

With regard to the exact age of the intrusions a difficulty arises through lack of direct evidence. With few exceptions the country-rock surrounding the intrusions is of Pre-Cambrian age. The Galloway granodiorites and adamellites in the Southern Uplands of Scotland are intrusive into Ordovician and Silurian rocks, so is the well-known Shap adamellite (2 square miles) in the English Lake District; while the largest single intrusion in the British Isles—the Leinster granite—is definitely Post-Silurian and Pre-Carboniferous in age. The small intrusions in the Manx Slates in the Isle of Man, comprising the Dhoon² porphyritic microgranite, and Foxdale granite, are also probably of Caledonian age. Now, had the granites been in place before the culmination of the Caledonian earth-movements they would undoubtedly show the effect of the intense pressures to which they would have been subjected. They do not do

¹ Richey, J. E., and Thomas, H. H., *Quar. Journ. Geol. Soc.*, **88** (1932), p. 787.

² Nockolds, S. R., "The Dhoon (Isle of Man) Granite," *Min. Mag.*, **22** (1931), p. 494.

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so, and hence must be younger than the acme of the revolution. In one or two cases the marginal portions of the granites show some foliation, and were therefore intruded before the earth-movements had entirely ceased. An upper limit to their age is fixed by the occurrence of boulders of the Kincardineshire granodiorite in the local base of the Lower Old Red Sandstone.¹ Not all of the intrusions, however, were of exactly this age, as some (e.g. the Ben Cruachan and Ben Nevis granodiorites) cut, and are therefore younger than, the Lower Old Red Sandstone lavas. The "Newer Intrusives" thus include two age-groups: an earlier Caledonian, and a later (Post-Lower, Pre-Middle Old Red Sandstone) series. Among others whose age is not definitely proven are the Mountsorrel Granite² and associated dioritic and gabbroic intrusions. In their petrographic characters they are distinctly Caledonian.

When the petrology of the intrusions is examined it is found that there are no marked differences between the two series, such as might be used as a criterion of relative age, and in many cases it is not possible to date the intrusions precisely.

Although the intrusions are dominantly acid, many other kinds of plutonic rock are associated with the great granite bosses, either as marginal facies or as satellitic intrusions situated near to the main masses. Chief among these are diorites and monzonites; though basic types, including gabbro and norite, are not uncommon, and ultrabasic rocks are sometimes found.

The "Newer Intrusives" of Scotland are all either adamellites or granodiorites, with still more basic peripheral modifications in some cases (tonalites and diorites).

The **South-East** and **Central Highlands** are remarkable not only for the number of great intrusions found in these parts of Scotland, but also the striking difference in their mode of occurrence: the irregularity and complexity of the forms assumed by the "Older (Pre-Torridonian) Granites" contrasts strongly with the comparative simplicity of form and uniformity of the "Newer Intrusives." Of these several masses, the Kincardineshire adamellite is the least variable; it contains the most plagioclase and is interesting on account of the apophyses thrown out on its south-western margin. These rapidly pass into porphyritic microgranites.

Some of the Caledonian granites increase in alkalinity towards the north or north-west, indicating intrusion at a time when the country was still under the influence of the Caledonian earth-movements, and suggesting the partial separation of the products of earlier from those of later crystallization, with the migration of the latter into regions of less stress. The same principle is illustrated

¹ "Summary of Progress," *Geol. Surv.* (1901), p. 111.

² Taylor, J. H., *Geol. Mag.* (1934), p. 1.

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by the occurrence of small intrusions of intermediate, basic and ultrabasic rocks near to the major acid intrusions. These are of earlier formation than the latter, and in several cases occur at the south-eastern side of the main masses. This is well illustrated by the Glen Tilt complex which comprises augite-diorite, tonalite, hornblende-adamellite, biotite-granite, muscovite-granite, and granite-aplite. A similar complex at Garabal Hill,¹ near the head of Loch Lomond, includes augite- and mica-diorites, norites and rocks bordering on hornblende-peridotites, now largely serpentinized, as well as tonalite and granite. This complex occurs along the south-eastern margin of the porphyritic granite of Glen Fyne.

It would be unprofitable to discuss the petrology of each of these intrusions, and only the more important points can be here considered. A constant feature is the occurrence of abundant plagioclase in addition to orthoclase in the more acid intrusions: most of the granites belonging to both age-groups are adamellites or granodiorites. This is illustrated by the so-called **Galloway** granites,² which are distinguished as the Criffel-Dalbeattie, Cairnsmore of Fleet and Loch Dee *massifs* (Fig. 135).

The largest of the "Galloway Granites" is the Criffell-Dalbeattie Complex³ which consists of an early quartz-diorite forming a separate intrusion adjacent to the main one. The latter is granodiorite including an earlier aphyric member, which is marginal to the later, central, porphyritic granodiorite. The Galloway granodiorites are associated with a well-developed dyke-swarm,⁴ some members of which trend in the same direction as the strike of the Silurian sedimentary rocks, while others are at right angles to this direction (N.W. to S.E. and N.E. to S.W.). The dyke rocks fall into three groups: a very small proportion (5 per cent) are lamprophyres, chiefly hornblending types (spessartites), but a few are kersantites. The chief members of the swarm, to the extent of 95 per cent, broadly compare with the plutonic types, though the majority have been recorded as "porphyrites" (intermediate medium-grained dyke-rocks of dioritic composition), but this is really a field term only. However, the porphyritic microdiorites are associated with, and grade into, microtonalites and ultimately microgranodiorites, both containing quartz, which in the latter is accompanied by subordinate potassic feldspar.

¹ Nockolds, S. R., "Garabal Hill-Glen Fyne Igneous Complex," *Quar. Journ. Geol. Soc.*, **96** (1940), p. 451.

² Non-British readers may note that Galloway is a name applied to part of south-western Scotland; but Galway is a county in Northern Ireland—both contain important granites.

³ Phillips, W. J., "The Criffell-Dalbeattie Igneous Complex," *Quar. Journ. Geol. Soc.*, **112** (1956), p. 221.

⁴ Phillips, W. J., "The Minor Intrusive Suite associated with the Criffell-Dalbeattie Granodiorite Complex," *Proc. Geol. Assoc.*, **67** (1956), p. 103.

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The Loch Dee Complex comprises a central outcrop of pale coloured biotite-granite, surrounded by biotite- and hornblende-tonalite. These tonalites are crowded with sedimentary xenoliths and pass into a strongly biotitic facies where they penetrate into the wall rock. Still more basic rocks are quartz-norites and norites

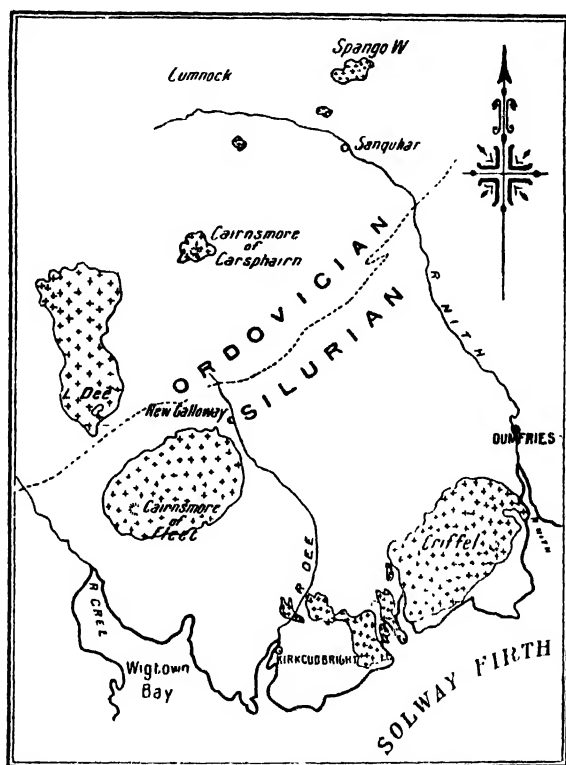


FIG. 135

Map of the Galloway district, showing the distribution of the granite complexes.
Scale: 1 inch = $13\frac{1}{2}$ miles.

which are marginal in position. A zone of hybrid rocks is believed to have resulted from admixture of the sub-magmas which gave rise to the tonalite and norite respectively.¹

The rock of the Cairnsmore of Fleet *massif* is somewhat less calcic, being typically a biotite-adamellite, grading into biotite-muscovite-granite. A clove-coloured sphenc distinguishes the Criffel tonalite, which is extensively quarried at Dalbeattie.

¹ Gardiner, C. J., and Reynolds, S. H., "The Loch Doon Granite Area, Galloway," *Quar. Journ. Geol. Soc.*, **88** (1932), p. 1.

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In addition to the *massifs* mentioned above, smaller intrusions of granodiorite associated with tonalite and quartz-norite, and evidently of the same age and origin, are found in the Galloway district at the Mull of Galloway, on the east side of the Cree, south of Creetown, at Cairnmore of Carsphairn,¹ Spango Water and south of New Cumnock (Fig. 135). There are also many dykes of Caledonoid trend of which the chief are porphyritic microdiorites, accompanied by a variety of acid and lamprophyric types.

The main structural and stratigraphical units established in Scotland extend across into Ireland where again Caledonian granites form important features in the Dalradian and Lower Palaeozoic tracts. The largest is the Leinster Granite; but the one which has been most exhaustively studied is the Donegal Granite illustrated in Fig. 93 and referred to in the discussion on the origin of granites. Similar granites occur also in the adjacent counties of Galway and Mayo.

The most alkaline of the Caledonian granites forms part of the **Leinster Granite** (625 square miles) in south-eastern Ireland. The main intrusion is an alkali-granite with an excess of potash over soda; but in some of the subsidiary intrusions, which are doubtless comagmatic with the main mass, soda predominates. Thus a soda-granite occurs at Aughrim, while another forms the summit of Croghan Kinshela and consists chiefly of a brilliant white sodic feldspar and grey quartz. In the north-eastern part of the complex three granites are distinguishable: the main type (central) is muscovite-granite, marginal to which is a porphyritic microcline-granite which occurs also in the roof-zone of the former. Finally the outer zone is finer-grained and granodioritic while the adjacent rocks external to the complex proper are migmatites. A noteworthy feature is a striking fluxional arrangement of dark schlieren and, particularly in the microcline-granite, of the porphyritic feldspars. Pegmatites and associated aplites occur within the complex.²

In most cases minor intrusions in the form of dykes and sills are closely associated with the "Newer Intrusives" and are considered below (p. 465). Mica-lamprophyres are less widely distributed, but occur in Galloway and in the Lake District, where they are connected with the Shap adamellite.

A second great group of "Newer Intrusives" includes the complexes of basic rock occurring in Aberdeenshire at Huntly, Haddo, Arnage, and Inch (Fig. 136). Although the various outcrops near Arnage are separated by a mantle of Highland Schists, they are probably

¹ Deer, W. A., "The Cairnmore of Carsphairn Igneous Complex," *Quar. Journ. Geol. Soc.*, **91** (1935), p. 47.

² Brindley, J. C., "The geology of the northern end of the Leinster Granite, Part I . . .," *Proc. Roy. Irish Acad.*, **56b** (1954), pp. 159-90.

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connected underground and are the exposed portions of an irregular sheet.¹ The igneous rocks are essentially norites. Between the norite and the country-rock is a wide zone of "contaminated norite," differing markedly from both intrusive and invaded rock. The contaminated rock is crowded with xenoliths in all stages of absorption,

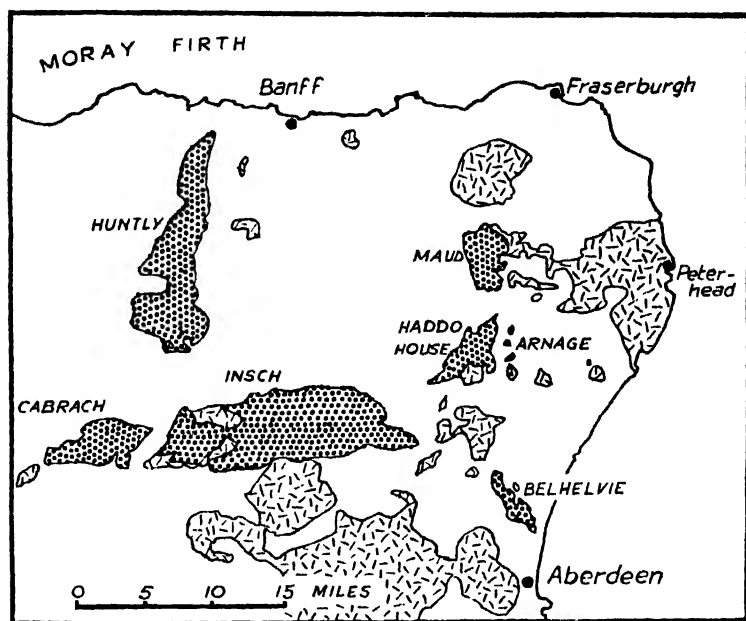


FIG. 136

Map of part of north-eastern Scotland, showing the distribution of the granitic and basic igneous complexes. (Based largely on the work of H. H. Read.)

and is usually rich in minerals, such as garnet, cordierite (Fig. 53) and spinel, not normally found in pure igneous rocks. It is noteworthy, however, that some of the slides of the contaminated rock do not appear to differ essentially from normal igneous rocks. Thus one type from Kinharrachie, which consists of hornblende, plagioclase and interstitial quartz, is for all practical purposes a quartz-hornblende-diorite.

The Huntly mass² is a sheeted complex some 50 square miles in area, within which a wonderful variety of rocks of special petrological interest are exposed. They include peridotites, olivine-gabbros, troctolites, norites, as well as small granite bosses. Some of the basic

¹ Read, H. H., "Petrology of the Arnage District," *Quar. Journ. Geol. Soc.*, 79 (1923), pp. 447-86.

² Read, H. H., *Geol. Mag.* (1924), p. 433, and "Geology of Banff, Huntly and Turriff," *Mem. Geol. Surv.* (1923).

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rocks exhibit a striking stratification, rivalling the famous banded gabbros in Skye.

The Belhelvie Complex (Fig. 136) is essentially similar as regards its rock-types, and is best known for the troctolites (Fig. 111) which are represented in most teaching collections.¹

DEVONIAN

Igneous rocks of Devonian age are important in **South Devon** and in **Cornwall**, and comprise basic and acid lavas (the latter being very subordinate to the former), together with tuffs and minor intrusions (Figs. 141 and 142). The extrusive rocks cover a large area in the neighbourhood of Totnes, particularly around the village of Ashprington, and are hence often referred to as the Ashprington Volcanic Series. From this locality they range southwards through Modbury, Saltash, Liskeard, and thence, swinging north of the Bodmin Moor Granite, reach the north coast of Cornwall near Padstow and Port Isaac. Within this tract the igneous rocks possess peculiarities indicating community of origin: the area is, in fact, a good illustration of a petrographic province, or rather, a portion of such, since the same series of Devonian lavas, tuffs and intrusions ranges across central Europe, through the Vosges and Harz Mountains, into Moravia and Nassau.

The original character of the rocks is best preserved in the northern part of the area, but towards the south the rocks become more and more affected by the Armorican earth-movements; indeed, in the Plymouth district they are so altered that their recognition as of igneous origin is difficult, while it is usually quite impossible to distinguish between lavas and intrusions.

In south-eastern Devonshire the eruptions commenced in early Middle Devonian times. The sites of the volcanoes have not been located, but it is clear that they must have been situated at a considerable distance from the shore-line of the Devonian Continent. Consequently the lavas are submarine, and are interbedded with limestone and other normal marine sediments. The basic lavas are spilites showing pillow-structure and the high degree of vesicularity characteristic of these rocks. These features are particularly well exhibited at Chipley, where a thickness of 70 feet of pillow-lava is exposed.²

In the Plymouth-Liskeard district the lavas are much decomposed, especially those belonging to the Middle Devonian. Those interbedded with the Upper Devonian sedimentary rocks are slightly

¹ Stewart, F., "The Gabbroic Complex of Belhelvie, Aberdeenshire," *Quar. Journ. Geol. Soc.*, **102** (1947), p. 465.

² "Geology of Newton Abbot," *Mem. Geol. Surv.* (1913), pp. 54-6.

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better preserved, and are seen to be chiefly spilites. Farther to the north, in the neighbourhood of Tavistock and Launceston, similar rocks occur in the Upper Devonian-Lower Culm.

On the north coast of Cornwall no igneous rocks are found in the Lower and Middle Devonian, but in the Frasnian division of the Upper Devonian they attain to their maximum development: at Pentire Point, near Padstow, the thickness of pillow-lava, well exposed in the sea-cliffs, exceeds 250 feet.¹

In addition to the abundant basic flows, others of acid composition occasionally occur in the neighbourhood of Newton Abbot and Ivybridge. Like the spilites with which they are associated, a high soda-content is characteristic, and the rocks may be referred to the sodic rhyolites (quartz-keratophyres of some authors). They are much brecciated, but it is difficult to decide whether this is the result of flow-movements in an extremely viscous lava, or of explosions in the vents.

The minor intrusions range in composition from basic to ultrabasic, and comprise three chief types; dolerites, minverites and picrites.

The dolerites are coarse-grained, non-vesicular rocks forming sills, in some of which large quarries have been opened for road metal, as at Trusham in the Newton Abbot district.² The augite is a normal brown variety, which in some specimens merges into titanaugite. No basic plagioclase is found in these rocks, but albite, some of which is primary, usually makes up nearly the whole of the feldspar present. An aplitic modification of the albite-dolerites is not uncommon. It occurs in veins and segregations, and consists essentially of micropertthitic feldspars.

The minverites are feebly developed in Devonshire, but are the dominant intrusive types in Cornwall, both in the neighbourhood of Plymouth and in the Padstow district. At the latter locality they form sills up to 70 feet in thickness. The type-specimens come from the Rock Quarry, St. Minver, on the Camel Estuary. In these rocks barkevikitic amphibole is the dominant coloured mineral, though it is sometimes accompanied by olivine, titanaugite and bronze-coloured biotite. As in the lavas, the feldspar is chiefly albite, but some anorthoclase occurs in addition. The texture varies considerably: some types are coarsely ophitic; in others there is a tendency towards idiomorphism of the dark minerals; while others again are compact, fine-grained and resemble some camptonites.

No fresh olivine has yet been found in the basic intrusions, but serpentinous pseudomorphs, embedded in the pyroxene, are not

¹ Dewey, H., *Proc. Geol. Assoc.*, 25 (1914), pp. 165-73.

² "Newton Abbot," *Mem. Geol. Surv.* (1913), pp. 59-63.

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uncommon in the dolerites. By increase in the amount of olivine, at the expense of the feldspar, there is a gradation to picrites, a few of which occur in this cycle. Teall has described the augite-picrite of Menheniot (Clicker Tor). Typical hornblende-picrite is found at Molenick in the Plymouth district, while augite-picrite occurs at Highweek in the Newton Abbot district.

Each of these intrusive types strongly resembles the contemporaneous rocks in the continuation of the province in central Europe.

THE OLD RED SANDSTONE LAVAS¹

Igneous activity during the Devonian Period was by no means restricted to the submarine eruptions that occurred in the area now occupied by Devon and Cornwall: in northern Britain and north-western Ireland contemporary volcanoes poured out an enormous quantity of lava and ash of quite different types from those described above, although of the same age. The Old Red Sandstone cycle is complete, comprising an extrusive phase, the dominant type being andesite; a phase of major (plutonic) intrusion, during which some of the largest granite masses in this country came into being; and a phase of minor intrusion, when porphyritic microdiorites were injected as dykes.

The lavas were poured out from volcanoes of the central type located along lines of instability connected with the Caledonian earth-movements. One series of volcanoes bordered the Midland Valley of Scotland on the north, and extended across into Ireland. Another series bordered it on the south (Fig. 137). Relics of the outpourings from the former occur in the Ochil and Sidlaw Hills, and the Lorne district of Argyllshire, including Ben Nevis and Glencoe; while the latter are represented by the lavas, ashes and intrusions of the Pentland Hills,² the Braid Hills, the Cheviot Hills (Fig. 139), and a small area in North Ayrshire.³

The volcanic rocks rest upon an uneven eroded land surface consisting of schists in the South-West Highlands, and folded Silurian rocks in the Cheviots and central Lowlands. Locally basal conglomerates and breccias are intercalated, and are of interest inasmuch as they contain pebbles of lava, thus proving the existence of even earlier flows. The basal sediments have yielded fragmentary

¹ This volcanic cycle is of Devonian age, but the extrusives are interstratified with rocks of "continental" facies—the Old Red Sandstone. These contrast strikingly with the marine facies of the Devonian, typically developed in Devon and Cornwall in the extreme south-west of England.

² Mykura, W., "The Lower Old Sandstone igneous rocks of the Pentland Hills," *Bull. Geol. Surv. Gt. Br.*, 16 (1960), pp. 135–55.

³ MacGregor, A. G., in "Geology of N. Ayrshire," *Mem. Geol. Surv.*, 1930, p. 28.

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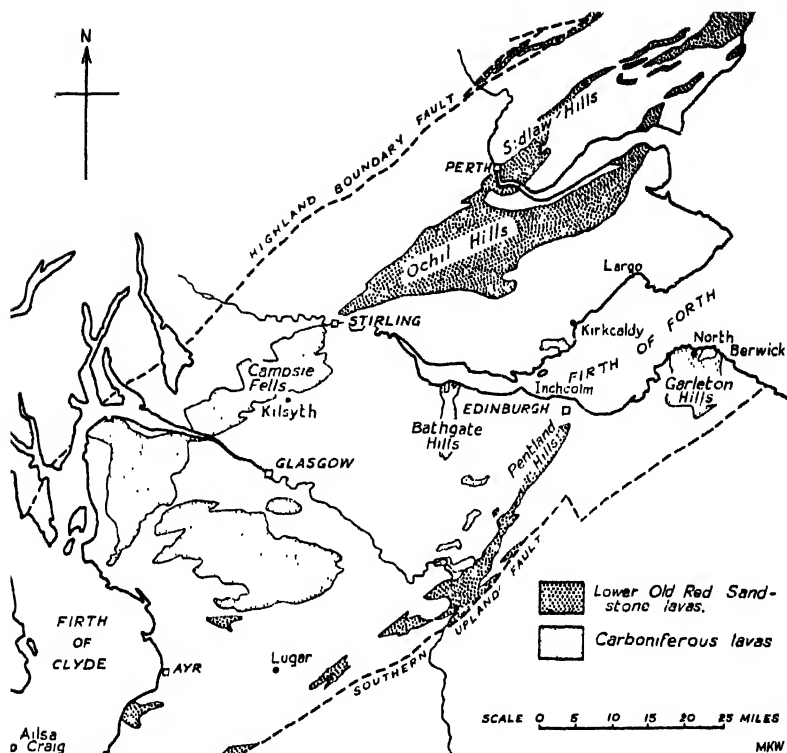


FIG. 137

Map of the Old Red Sandstone and Carboniferous lavas of the Midland Valley of Scotland. (Based on maps of the Geological Survey.)

plant remains (*Psilophyton* and *Pachytheca devonica*)¹ in Lorne and Glencoe, which prove the age of the overlying volcanic rocks as Lower Old Red Sandstone. Further, they are succeeded unconformably by Upper Old Red Sandstone in some localities.

The lavas themselves are frequently brecciated (block-lavas) and in some cases can only with difficulty be distinguished from interbedded agglomerates. The proportion of pyroclasts to lavas is small, as the eruptions were not of a violently explosive type. It is probable that the lavas were in part erupted under subaerial conditions: the upper portions of the flows are frequently reddened, due to atmospheric weathering between one eruption and the next. Occasional shales and sandstones occur interbedded with the lavas, and sand-filled crevices and "sandstone dykes" are characteristic.

Although the succession has been the subject of careful study at

¹ "Geology of Ben Nevis and Glencoe," *Mem. Geol. Surv.* (1916), p. 89.

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many localities, it is possible to correlate individual flows over restricted areas only, because the lavas erupted from neighbouring centres differ in composition, although the volcanoes were so close to one another that their products interdigitate. The dominant type of lava is a basic andesite (verging on basalt), and this is interbedded with rhyolite, trachyte and basalt. The succession established at Glencoe may be regarded as typical. This comprises the following:

	Feet
7. Andesites and rhyolites	about 300
6. Shales and grits	,, 50
5. Rhyolite	,, 250
4. Hornblende-andesites	,, 900
3. Agglomerate and shale	,, 250
2. Rhyolites	,, 450
1. Augite-andesites	,, 1,500

The more basic lavas contain small red pseudomorphs after olivine and bear a close resemblance to mugearites. In others olivine is less abundant and is accompanied by pale green augite. In the typical augite-andesites only the latter occurs as phenocrysts. Basalts and basic andesites with phenocrysts of feldspar are uncommon in the west (Glencoe and Ben Nevis), but a very distinctive type occurs near the top of the succession in the Pentlands. This rock is distinguished as the "Carnethy porphyry," and is particularly rich in fluxionally arranged, platy phenocrysts of plagioclase. One of the basic lavas in Glencoe contains the rare red variety of epidote, withamite, occurring as blood-red crystals in druses.

The less basic lavas include enstatite-, hornblende- and mica-andesites. In addition to phenocrysts of coloured minerals, these andesites are rich in porphyritic feldspars. The Cheviot lavas¹ total some 1,200 feet in thickness, and include dominant augite-hypersthene-andesites, glassy andesites, trachyandesites and curious types termed by the Survey Officers "oligoclase-trachytes." These we would classify as oligoclase leuco-andesites.

The Old Red Sandstone lavas which build the Sidlaw Hills in eastern Scotland are fresher than those in most other areas where volcanic rocks of this age occur. They include basalts, basaltic andesites and true andesites in variety, together with rare trachy-andesites and comagmatic minor intrusions.^{2,3}

¹ Carruthers, R. G., and others, "Geology of the Cheviot Hills," *Mem. Geol. Surv.* (1932), p. 8.

² Harry, W. T., "The Old Red Sandstone lavas of the western Sidlaw Hills, Perthshire," *Geol. Mag.*, 93 (1956), pp. 43-56.

³ Harry, W. T., "Old Red Sandstone Lavas of the Eastern Sidlaws," *Trans. Edin. Geol. Soc.*, 17 (1958), pp. 105-12.

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In chemical composition the lavas noted above are evidently closely related to the intrusive augite-diorites, kentallenites and monzonites of the plutonic phase. Some of the lavas are rich in orthoclase, particularly certain flows from Lorne, and consequently are to be regarded as trachybasalts or trachyandesites, according to their degree of basicity.

The dacites connect the acid lavas with the hornblende- and mica-andesites, from which they differ chiefly in the occurrence of phenocrysts of quartz. They are so closely similar in appearance to many of the rhyolites, that in some districts the two types of rock have not been differentiated in the field.

The rhyolites are chemically related to the most acid granites and microgranites of the succeeding phases, but vary considerably in texture among themselves. Some show beautiful flow-structure, others are cryptocrystalline through devitrification of an originally glassy rock, while many are spherulitic. The phenocrysts include albite, orthoclase, biotite and quartz. It is usually the case that the more acid andesites grade into rhyolites, but in the Pentland Hills they are associated with potassic trachytes.

THE OLD RED SANDSTONE MAJOR INTRUSIONS

Reference has already been made to the general characters of the post-Lower Old Red Sandstone major intrusives. They range in composition from acid granites to basic rocks rich in olivine, while locally feldspar-free ultrabasic types occur. Outcrops of acid rocks are far more extensive than those of intermediate and basic composition, and in volume the former are in large excess over the latter, *i.e.*, the magma, as represented by the plutonic rocks, was essentially acid. Direct evidence of the age of these intrusions is sometimes forthcoming. Thus on Ben Nevis, in Glencoe and the Cheviot Hills, the granites cut the Old Red Sandstone lavas, and must therefore be younger than the latter. In other cases, however, such direct evidence is wanting, and the age of many of the granite masses in the Highlands is therefore largely a matter of speculation. In many cases it is not possible to make a more precise statement than "post-Silurian, pre-Middle Old Red Sandstone." Facts of general interest in connection with the Old Red Sandstone major intrusions are: their large size, their composite nature, the occurrence of satellitic intrusions of basic to ultra-basic composition and the sharp, regular character of the highly inclined junctions with the country-rock. In these several particulars, the intrusions differ from the so-called "Older Granites" of Scotland.

With regard to the form of the major intrusions, it is clear that

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many are true bosses: the outcrops are in many cases roughly circular, the contacts are steeply inclined, while the manner in which they cut across the bedding and foliation planes of the country-rock is strikingly shown on the geological maps. On account of the marked similarity of composition between rocks of this age over very extended areas, it may be suspected that they are connected underground, and that the bosses are cupolas rising from the tops of a batholith. The intrusions occurring at Ben Nevis and Glencoe are of rather special interest as they bear evidence of an unusual method of intrusion.

The granites of Glencoe form part of one of the greatest intrusive masses in Scotland, known as the Etive Complex¹ (Fig. 138). The earliest and main portion of this complex is the Moor of Rannoch Granite; this is cut by the Ben Cruachan Granite, and the latter again by the Starav boss. The evidence seems clear that the dyke-phase, referred to below, supervened between the injection of the Cruachan and Starav masses. We thus have an interesting case of one granite boss penetrating another of somewhat earlier date. It is believed that the second injection was consequent upon the down-faulting of a cylindrical plug of country-rock (in this case the Cruachan Granite). It has already been shown that cylindrical or conical fractures may be produced above the top of an advancing plug of magma. In at least three cases in the area under consideration, such fracturing was followed by collapse or foundering into the magma underlying the tract surrounded by the fault. Such phenomena have been termed cauldron subsidences.² In Glencoe and on Ben Nevis the lavas owe their preservation to such "piston-faulting" (Fig. 63). In the latter locality, the igneous *massif* consists of three approximately concentric zones: the outer granite, the inner granite and, in the centre, the outcrop, roughly circular in shape, of the lavas resting on a floor of schist. As in the Etive Complex the dyke-phase intervened between the injection of the outer and the inner granites. In the light of the clear evidence obtained at Glencoe, it is probable that cauldron subsidence operated on three occasions at Ben Nevis, while the outer granite itself consists of three fractions forming arcuate outcrops.³

With regard to the petrography of the major intrusions the dominant types are adamellites and granodiorites,⁴ *i.e.* a large

¹ Anderson, J. G. C., "The Etive Complex," *Quar. Journ. Geol. Soc.*, **93** (1937), p. 487.

² Clough, C. T., Maufe, H. B., Bailey, E. B., *Quar. Journ. Geol. Soc.*, **65** (1909), p. 611.

³ Anderson, J. G. C., "The Marginal Intrusions of Ben Nevis," *Trans. Geol. Soc., Glasgow*, **19** (1935), p. 225.

⁴ The student should note that in the "Geology of Ben Nevis and Glencoe," *Mem. Geol. Surv.* (1916), the rock-names are not used in exactly the same sense as in this textbook.

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porphyritic. The Cruachan adamellite consists of cryptoperthite, zoned plagioclase and quartz. The dominant coloured constituent is green hornblende, but this is accompanied by biotite, either intergrown with or moulded upon the amphibole. The Starav adamellite is still more acid, the two feldspars are present in approximately equal proportions, but coloured minerals are less abundant. More acid still is a high level modification of the Cruachan mass, in which the coloured mineral (biotite) is very subordinate, the rock consisting of quartz and alkali-feldspars (chiefly perthites). Leucogranite of this type seems to be widely distributed among the "Newer Intrusives."

The Cheviot granite¹ (Fig. 139) is a partly unroofed laccolith, the exposed area being 22 square miles. It is noteworthy on account of the modification it has undergone through assimilation of the country rock. The normal rock is a pink micrographic granite deficient in coloured minerals; but in places, particularly where the magma has penetrated *lit-par-lit* into the surrounding lavas, it is highly contaminated, of dioritic aspect and locally contains augite.

In the granophyric type biotite is the only coloured silicate; but in the marginal facies biotite is accompanied by pyroxene, both diopsidic augite and hypersthene being present; quartz may amount to a few per cent only, and may be absent. Thus these marginal rocks are, in fact, monzonites and diorites, obviously derived, by contamination, from granitic magma.

The more basic intrusions form small bosses, as a rule not far removed from the great granite intrusions described above. We shall again largely restrict our attention to the South-West Highlands, where the age of the intrusions is not questioned. The chief types of intermediate, basic and ultrabasic rocks occurring in Argyllshire and the adjacent islands are shown in tabular form below.

<i>Intermediate</i>	<i>Basic</i>	<i>Ultrabasic</i>
Augite-diorite		
Hornblende-augite-diorite	Kentallenite	Augite-picrite
Monzonite		
Hornblende-diorite	Hornblende-	Hornblendite
Appinite	gabbro	Hornblende-peridotite

The dominant rock types are those in which pale green augite (diopside or malacolite) is the characteristic coloured constituent:

¹ Carruthers, R. G., and Others, "Geology of the Cheviot Hills," *Mem. Geol. Surv.* (1932), p. 87. Thomas, H. H., in "Geology of the Cheviot Hills," *Mem. Geol. Surv.* (1932), p. 90; Jhingran, A. G., "The Cheviot Granite, etc.," *Quar. Journ. Geol. Soc.*, **98** (1943), p. 241.

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normal hornblende-syenites and diorites are rare among the intermediate rocks; but many of the meladiorites are rich in hornblende and have been termed "appinites."¹ With decreasing feldspar these rocks grade into hornblendite, and with the incoming of olivine into hornblende-peridotite of the Cortlandt type. In Colonsay, one of the Hebrides adjacent to the mainland, a similar series of intrusions has been described by the Survey.² Direct evidence of age is wanting, but on the ground of petrographic similarity they are referred to the Old Red Sandstone. Since, in some cases, they are intrusive into breccia or agglomerate it has been suggested that they mark the site of explosion-vents and are themselves vent-intrusions.

THE LOWER OLD RED SANDSTONE DYKE-PHASE

One of the most striking features of this igneous cycle is the well-developed dyke-phase which in general followed, though to some extent it overlapped, the plutonic phase. The dyke-phase reaches its maximum development in western Argyllshire, where the uprise of magma into a series of closely spaced parallel fractures formed the **Etive swarm** of N.E.-S.W. dykes (Fig. 138). The dykes are restricted to a comparatively narrow belt of country very much elongated in the direction of the dykes themselves. There can be little doubt that this tract was located above the body of magma, the comparatively thin roof of which constituted a belt of weakness in the crust. The magma must have been closely similar in composition to the lavas poured out during the extrusive phase, since the dominant type among the dyke-rocks is porphyritic microdiorite together with aphyric types.³ In the former rocks idiomorphic crystals of hornblende and biotite, together with plagioclase ranging from oligoclase to andesine, occur as phenocrysts. The same pale green augite which characterizes the more basic plutonic intrusions is occasionally found in the dykes, and is sometimes accompanied by orthorhombic pyroxene. In addition to the microdiorites more acid, as well as more basic types are found. The former include porphyritic microgranites, with phenocrysts of quartz and alkali-feldspar, set in a variable groundmass. The proportion of coloured minerals in these rocks is low, and they are evidently closely related to the leucogranites. Acid dykes without phenocrysts and deficient in dark minerals are not uncommon; it seems reasonable to regard them as aplites complementary to the lamprophyres which also occur. The latter are of two different ages:

¹ "Ben Nevis and Glencoe," *Mem. Geol. Surv.* (1916), p. 168.

² "Geology of Colonsay and Oronsay," *Mem. Geol. Surv.* (1911), pp. 28-37.

³ Termed "malchite" in the Ben Nevis and Glencoe Memoir.

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some occurring as horizontal sheets are probably the earliest intrusions in the cycle, while others are clearly members of the Etive and Ben Nevis swarms. As might be expected, the dominant types are rich in hornblende (hornblende-lamprophyres), sometimes with olivine in addition (olivine-hornblende-lamprophyres), while mica-lamprophyres with augite and olivine are but feebly represented.

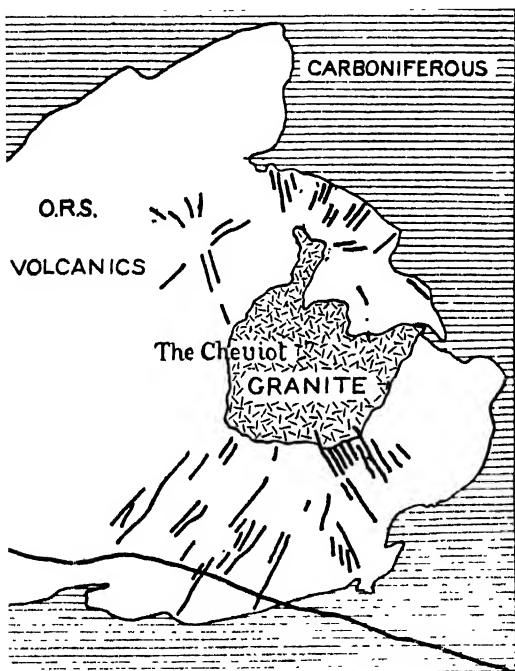


FIG. 139

Sketch-map to illustrate the Old Red Sandstone igneous cycle in the Cheviot Hills. Dykes are shown by thick lines. (*After H. Kynaston.*)

The dyke-phase in the Cheviot Hills includes two swarms differing slightly in age, but widely in trend. They freely cut the lavas, but few can be traced into the granite, which appears to be later than the majority of the dykes. The chief type is almost identical with the dominant type of lava, being augite-hypersthene-microdiorite or andesite, according to grain-size.

In conclusion, the igneous rocks of Lower Old Red Sandstone age in Scotland and the Border Country may be regarded as an excellent example of an igneous cycle. The magma was essentially calc-alkaline in facies and, although the three phases were distinct, the

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products are so closely related in composition as to leave no room for doubt as to their common origin. The contrast between the Devonian igneous rocks of South-West England and the Old Red Sandstone igneous rocks of Scotland and the Borders is to be accounted for in part by differences in the conditions of outpouring (the former being submarine and the latter subaerial), and in part by association with earth-movements of different kinds. On the one hand, the calc-alkaline character of the rocks is a result of the close connection between the Caledonian movements and the up-rise of the magma that formed the Old Red Sandstone igneous rocks. On the other, the Devonian area (Devon and Cornwall) far to the south was unaffected by the Caledonian earth movements; it was an area of regional subsidence, *i.e.* geosynclinal, and this led to the development of a spilitic suite, comprising spilitic pillow lavas, albite-trachytes (keratophyres) and associated albitic intrusives.

The main igneous cycle ceased in Lower Old Sandstone times, but there were feeble revivals of activity in Middle and Upper O.R.S. times. The andesitic lava and associated tuffs at Rhynie in Aberdeenshire are interbedded with Middle O.R.S. rocks. The tuffs are unique as they contain silicified plant-remains which have proved invaluable to palaeobotanists. In the Orkney Islands lavas and pyroclasts occur both in the Middle and the Upper O.R.S.

In the region of the Orkneys, north of the Scottish mainland, an explosive eruption of great violence brought to a close the period of folding and erosion that immediately preceded Upper Old Red Sandstone times. The products of the explosion were spread as a thick tuff over a wide area, and this is succeeded by a flow of olivine-basalt in the island of Hoy. Later still, the rocks, up to and including the Upper Old Red Sandstone, were cut by numerous dykes which "form a petrological group of remarkably interesting characters" that cannot be exactly matched anywhere else in Britain.¹ The dyke-rocks include leucocratic types, but the majority are melanocratic and thoroughly basic. The most interesting of the former are microsyenites that are very typical examples of their kind: normally highly feldspathic, one of them contains the highest percentage (nearly 11 per cent) of potash of all British analysed rocks, and consists almost exclusively of orthoclase. The melanocratic group includes three chief types: camptonites, three out of every four dykes falling in this category; monchiquites, to the extent of nearly a quarter of the whole number; and a few intrusive olivine-basalts. The camptonites may be regarded as the central type. They contain small olivines; zoned augite with green (perhaps chrome-diopside) cores surrounded by mantles of titanite, is characteristic.

¹ Flett, Sir John, in "Geology of the Orkneys," *Mem. Geol. Surv. Scotland*, (1935), p. 173.

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A third mafic component is basaltic hornblende. With the incoming of porphyritic plagioclase and the elimination of the basaltic hornblende the camptonites grade into basalts; while in the opposite direction, with the elimination of feldspar, they pass into monchiquites. In these the two essential constituents are olivine and augite, the latter making up two-thirds of the rock, embedded in brown glass, not in analcite, which, however, occurs in steam cavities and ocelli. Nepheline may occur as small crystals in the groundmass or as micropoikilitic patches. Finally, some varieties are rich in megascopic biotite and approach closely to alnöite. Comparison of the available analyses shows a fairly close correspondence between these monchiquites of the Orkneys and "nephelinites," suggesting that the former are the dyke-equivalents of the latter; but the monchiquites are rather poorer in silica, alumina and alkalies, thus stressing their melanocratic character and lamprophyric facies.

The age of these dykes is uncertain: they differ in type from the Old Red Sandstone dykes of other parts of Scotland, and may be either Carboniferous, Permian or Tertiary. In this connection it is significant that both camptonites and monchiquites are recorded from among the Permian dyke-rocks of Ayrshire, while one of the Permian vent-agglomerates in the same area has yielded blocks of monchiquite containing xenocrysts of anorthoclase.¹ At least one of the Orkney monchiquites also contains these xenocrysts, a fact which, taking into account the extreme rarity of such rocks, affords strong evidence that both occurrences are of the same age, *i.e.* Permian.

In retrospect it may be well to emphasize the close similarity, in all respects but scale, between the products of Old Red Sandstone igneous activity in northern Britain, and those occurring in the western mountain ranges of North and South America. Both cycles are of the Orogenic Continental type, and there can be little doubt that the parental magma was the same in both cases, and essentially andesitic (see discussion on origin of andesites, especially p. 298).

NOTE.—Readers may profitably consult a paper received when this chapter was in proof, entitled "Aspects of the Caledonian Magmatism," by H. H. Read, in *Liverpool and Manchester Geol. Journ.*, 2, pt. 4 (1961), pp. 653–83.

¹ Eyles, V. A., in "Geology of North Ayrshire," *Mem. Geol. Surv. Scotland*, (1930), p. 288.

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CARBONIFEROUS

DURING the Carboniferous Period Scotland was again the scene of widespread and long-continued igneous activity. In England contemporaneous eruptions led to the accumulation of tuffs and lavas in Derbyshire and the Isle of Man; while in Devon and Cornwall, as we have already seen, the period of activity which commenced in Middle Devonian times persisted into the early part of the Carboniferous. In each of these localities the cycle was limited to the extrusive phase and the phase of minor intrusions. There are no plutonic rocks comparable in size with the Caledonian and Old Red Sandstone bosses of granite.

(1) SCOTLAND

It is significant that the unstable area of the **Midland Valley of Scotland** was the site of most of the volcanoes of Lower Carboniferous age. The earliest eruptions occurred in the east, in the neighbourhood of North Berwick, and, as is usually the case, the lowest volcanic rock is a typical explosion-tuff. This is succeeded by a considerable thickness of basalt-flows, which, as a result of subsequent movement, are inclined towards the west, in which direction the lavas become progressively more acid, indicating an increase in the acidity of the magma as time went on. Thus, in the Garleton Hills the basalts are succeeded by thick sheets of trachyte. Farther to the west, in Midlothian, lavas of the same age (Calcareous Sandstone) occur as outliers, as at Arthur's Seat, in Edinburgh, where they rest on the Cementstone Group; and on the north bank of the Firth of Forth between Burntisland and Kirkcaldy. In the latter localities the eruptions were slightly later in date, belonging to the higher part of the Calcareous Sandstone, and extending up to the base of the Carboniferous Limestone.¹ Still farther west, extensive outcrops of Carboniferous igneous rocks occur which are relics of

¹ The term "Carboniferous Limestone" does not have the same age-significance in Scottish and in English stratigraphy: in Scotland the term covers only the limestone-bearing part of the Lower Carboniferous below which come the Oil Shale Group and the Cementstone Group. In England the term covers the *whole* of the Lower Carboniferous (the Avonian).

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a widespread plateau of basalt, the Clyde Plateau of A. Geikie, and, judging from existing outcrops, it must have once stretched continuously from beyond Arran in the west to Stirling in the east, and from the Highland Border southwards well into Ayrshire. A moderate estimate of the original area of the Clyde basalt-plateau is 2,500 square miles. In the Glasgow district the lavas form the terraced scarps of the Campsie Fells in the north, the escarpments of the Kilpatrick Hills in the north-west, and the Cathkin Hills in the south. Petrographically these basalts are of the same types as those of the more easterly areas. Several volcanic necks, some of considerable size, have been located. It is probable that the stacks of North Berwick Law and the Bass Rock mark the sites of vents; while Arthur's Seat, in Edinburgh, is a composite vent, built partly of agglomerate and partly of vent-intrusions, which are lithologically identical with the associated flows¹ (Figs. 54 and 55). Such is the well-known Lion's Haunch (intrusive) basalt. Vents are also of common occurrence in the Clyde Plateau. Some are choked with basalt, while others are filled with agglomerate. The vent known as Meikle Bin in the Glasgow district is important, since the associated intrusions include (intrusive) phonolitic trachytes.

In the eastern part of the Midland Valley the volcanoes soon became extinct; but those occurring farther to the west persisted for a much longer period, and new vents were established, indicating a progressive migration of the centres of eruption to the west. Thus in the Bathgate Hills, in the neighbourhood of Linlithgow and Bo'ness, eruptions, which commenced late in Calciferous Sandstone times, persisted into the Lower Limestone Group of the so-called Carboniferous Limestone of Scotland;² while at Bo'ness other flows are found in the Upper Limestone Group. This is also the case in the Saline Hills north of the Forth. Still later in date, in the Limestone Coal Group and Upper Limestone Group tuffs are interstratified with normal sediments at Dalry, North Ayrshire. Highly decomposed olivine-basalts, chiefly of Dalmeny type, occur in the Millstone Grit of the West of Scotland. In Ayrshire³ these volcanic rocks reach a maximum thickness of some 500 feet at Troon; they extend northwards to Stranraer, westwards into Arran, and possibly across into Ireland in the Ballycastle coalfield. Their highly decomposed condition is due to contemporaneous weathering effected largely by acid water formed by rotting of the luxurious vegetation which clothed the area at that time. The basalts in some cases have been converted into bauxitic clays.

¹ Clark, R. H., "Petrological Study of the Arthur's Seat Volcano," *Trans. Roy. Soc. Edin.*, 63 (1958), pp. 37-70.

² Falconer, J. D., *Trans. Roy. Soc. Edin.*, Vol. 45 (1906), part i, p. 139.

³ Wilson, G. V., and MacGregor, A. G., in "Geology of North Ayrshire," *Mem. Geol. Surv.* (1930), pp. 206 and 221.

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A much more important recrudescence of activity took place in Ayrshire in late Carboniferous to early Permian times. The lavas belonging to this subsidiary cycle occur in the Mauchline district, and are characterized by being more definitely alkaline in facies than the earlier rocks: they include types containing analcite and nepheline.¹

An outlier of the volcanic region of the Midland Valley of Scotland occurs to the south of the Southern Uplands extending for some fifty miles from the Kirkcudbright coast to the Langholm area lying north of the Canonbie coalfield. The lavas are basalts of various types, closely matching those of the main outcrop to the north. They rest directly and unconformably on the Old Red Sandstone red marls and reach a maximum thickness of 250 feet. Some 1,500 feet higher, near the top of the Cementstone Group and again above the latter in the lower part of the Oil Shale Group, basaltic volcanic rocks occur and doubtless originally linked up with their equivalents in the Midland Valley.²

The only other important tract in Scotland where igneous rocks of Carboniferous age occur is near the Anglo-Scottish border, in the **Tweed Valley**. In this case, also, the present outcrop is a mere fragment of a basalt plateau, which has disappeared under the influence of erosion. To the west of the present outcrop numerous plugs of basalt and agglomerate probably also belong to the Carboniferous Sandstone cycle. This is the case with the interesting intrusions in the Eildon Hills, near Melrose, referred to below.

The Basic Lavas

The Carboniferous lavas of Scotland have been exhaustively studied by Hatch, Watts, Tyrrell,³ A. G. MacGrégor,⁴ and S. I. Tomkeieff,⁵ and detailed classifications have been evolved, based chiefly upon mineralogical composition. The characteristic minerals of the Lower Carboniferous basalts are plagioclase, ranging in composition from oligoclase to bytownite, but labradorite is typical; a pale brown augite, occasionally with a tinge of mauve, though not with the full mauve tint of titanaugite; and olivine partly altered into serpentine or iddingsite. The accessories include orthoclase, analcite, biotite, hornblende, ilmenite and apatite. The three essential minerals are present as phenocrysts in widely varying proportions.

¹ Tyrrell, G. W., *Geol. Mag.*, 1912, pp. 69-80 and 120-31.

² Elliott, R. B., "Carboniferous Volcanic Rocks of the Langholm District," *Proc. Geol. Assoc.*, 71 (1960), p. 1.

³ Tyrrell, G. W., *Geol. Mag.* (1912), pp. 69-80 and 120-31.

⁴ "Classification of the Scottish Carboniferous Olivine-Basalts, etc.," *Trans. Geol. Soc. Glasgow*, 18 (1928), p. 324; and "Problems of Carboniferous-Permian volcanicity in Scotland," *Quar. Journ. Geol. Soc.*, 104 (1948), p. 133.

⁵ Tomkeieff, S. I., "Petro-chemistry of the Scottish Carboniferous-Permian Igneous Rocks," *Bull. Volcan., Série 2*, 1, pp. 61-87.

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The most distinctive megaphyric types have been described in the section dealing with the classification of basalts; but others also occur which are mineralogically of the same composition but contain only microphenocrysts (the Dalmeny, Hillhouse and Jedburgh types). Other basalts share the characters of two of the named types and may conveniently be referred to as, for example, Jedburgh-Dalmeny basalts, though this is a degree of nicety of interest only to specialists in these rocks.

It should be noted that these type-names are not restricted to lava-flows, but are applied also to small intrusive masses forming volcanic necks and sheets. Further, they represent arbitrarily chosen points in a continuously variable series.

In addition to true olivine-basalts typical mugearites occur as occasional intercalations among the olivine-basalts. As we have seen, mugearites are as basic as olivine-basalts, they contain olivine, and small amounts of biotite and a brown amphibole, but the plagioclase is oligoclase.

Most of these types are widely distributed in the Midland Valley, but there appears generally to be a greater development of the more feldspathic types in the west than in the east. Thus, on the Little Cumbrae, the lavas are Markle basalts and mugearites, while varieties richer in phenocrysts of coloured minerals are restricted to the numerous basaltic plugs and north-east or east-north-east dykes. In South Bute, however, the Markle, Jedburgh, Dunsapie and Craiglockhart types as well as mugearites are represented among the flows.¹

The Permian lavas of the Mauchline district of Ayrshire tend to be more basic, and, at the same time, more alkaline than the Lower Carboniferous basalts. They were emitted from sixty or more small vents which are lined with the materials forming the flows and choked with agglomerate. The lavas are largely olivine-basalts of the Hill house and Dalmeny types, but include also analcite-basanites, nepheline-basanites and limburgites. On the other hand the more felsic types, trachytes, trachyandesites, etc., are absent from the Permian lavas.²

The Intermediate Lavas

The less basic lavas, which are restricted to the main Lower Carboniferous cycle, are trachytes of various kinds. Some containing accessory nepheline are phonolitic trachytes. The intrusions associated with these flows are of the same composition and texture as

¹ Tyrrell, G. W., "Igneous Geology of the Cumbrae Islands," *Trans. Geol. Soc. Glasgow*, 16, part ii (1916-17), pp. 244-74.

² Tyrrell, G. W., *Trans. Geol. Soc. Glasgow*, 18, part ii (1927-28), p. 259; also Eyles, V. A., MacGregor, A. G., "Igneous Geology of Central Ayrshire," *ibid.*, 18, part ii (1930).

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the latter, and in many cases it is impossible to state definitely whether the rock is a flow or an intrusion.

The phonolitic trachytes consist essentially of sanidine, with which is usually associated a little plagioclase, and a green and slightly pleochroic soda-augite, together with some nepheline. Such rocks form the laccolith of Traprain Law, the sill of Hairy Craig, and the stocks of North Berwick Law and the Bass Rock. The Traprain Law rock contains a little sodalite in addition to analcite and nepheline, while the micro-syenite of the Bass Rock carries fayalite and nepheline.¹ In one case nepheline is sufficiently abundant to justify the use of the term phonolite. There is one exposure of this rock near Fintry, in the Campsie Fells.

A. G. MacGregor has described the Calciforous Sandstone lavas and associated intrusions from the fringe of the Clyde Plateau in North Ayrshire, and remarks that the occurrence of phonolitic trachytes here, as in the Garleton Hills farther east, emphasizes the alkaline affinities of the whole province. In this region also there is a unique nepheline-basanite in the same series, while the normal suite ranging from basalt to trachyte goes farther, yielding rhyolites.²

Trachytes are not uncommon in the central part of the Campsie Fells, round the Meikle Bin vent, where they form small elongated plugs or short dykes.³ Their state of preservation does not allow of detailed description, but it is claimed that there is here even a greater variety of types than is to be found among the trachytes of the Garleton Hills, East Lothian. Some contain "moss-like" pseudomorphs, which may well have been riebeckite originally. The latter mineral is found in the trachytes that occur as intrusions in the Eildon Hills. In some of the trachytes there is accessory quartz. In others large phenocrysts of sanidine are prominent, as at Peppercraig: these are perhaps best developed in certain rocks that are evidently links connecting the basalts with the true trachytes, *i.e.* they are trachybasalts (of the Banak type), resembling closely the Yellowstone Park type-rock, in that plagioclase as basic as labradorite is associated with much sanidine. The latter occurs as a peripheral zone round the former in the groundmass, and locally as phenocrysts two inches in diameter, enclosing smaller plagioclase crystals.

The Intrusions

As noted above, there are no plutonic intrusions belonging to this cycle, but the occurrence of ultrabasic plutonic rocks among the

¹ Campbell, R., *Trans. Edin. Geol. Soc.*, 13 (1933), p. 126.

² "Geology of North Ayrshire," *Mem. Geol. Surv.* (1930), p. 89.

³ Bailey, E. B., in "Geology of Glasgow District," 2nd Edn., *Mem. Geol. Surv.* (1925), p. 185.

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constituents of the agglomerates filling the vents and as numerous xenoliths in the basic lavas, points to the deep-seated presence of such rocks as dunite, augite-peridotite, pyroxenite and biotitite, none of which has as yet been exposed by denudation.

Minor intrusions are, however, numerous and varied in composition and habit. As the main eruptions were followed by revivals of activity at three later periods: in the Carboniferous Limestone, in the Millstone Grit, and in the Permian, it is difficult to assign a definite age to any particular intrusion. There is no clearly defined dyke-phase restricted to a relatively short interval of time; on the contrary the injection of the minor intrusions was spread over the whole of the cycle: the sills that are petrologically similar to the lavas with which they are interbedded were probably contemporaneous with these flows—the vent-intrusions and plugs certainly were. There can be little doubt that the trachytic plugs and other intrusions in the Midland Valley and in the Eildon Hills were contemporaneous with the upper lavas, of Cementstone age, in the Garleton Hills. Although the minor intrusions in the eastern part of the Midland Valley do not occur at higher horizons than the Carboniferous Limestone, in the western part of this tract they range as high as the red (?) Permian Sandstones in Ayrshire. In mineral composition these intrusions are clearly related to the Mauchline lavas, and were injected at a period somewhat later than the extrusion of the Permian lavas. They occur in the form of sills reaching 200 feet in thickness; and although they have a moderately wide range of composition, they are evidently comagmatic and consist of different proportions of the following minerals: olivine, titanaugite, labradorite, ilmenite and analcite, with nepheline, barkevikite and other minerals as accessories. The occurrence of primary analcite in many of these intrusions has been satisfactorily established. Its presence in the magma undoubtedly decreased the viscosity, and hence allowed an unusual degree of differentiation, which is demonstrated by the stratiform nature of many of the larger intrusions. No intrusion in this cycle shows this more clearly than the Lugar Sill in Ayrshire,¹ which includes some half-dozen distinct rock-types, *i.e.*, teschenite, theralite, picrite, peridotite and lugarite.

Two series have been distinguished among these minor intrusions, one characterized by analcite and the other by nepheline. The former series includes analcite-syenite, a typical example occurring at Howford Bridge, Ayrshire, teschenite and crinanite. The most distinctive members of the second group are theralite and lugarite. There are intermediate types, containing both analcite and nepheline, such as the well-known essexites of Crawfordjohn and Lennoxton. Localized intrusions of basic lamprophyric rocks

¹ Tyrrell, G. W., *Trans. Geol. Soc. Glasgow*, 21 (1948), p. 15.

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including camptonites and monchiquites occur in or near the Ayrshire vents.¹

The most typical and the most widespread of these rock-types is teschenite, which is common in the Glasgow district, in Ayrshire, in East Lothian and Fifeshire. Augite-picrite is frequently associated with the teschenites, of which it is an ultrabasic differentiate. This is the case, for example, in the Inchcolm sill which forms an island in the Firth of Forth.² Typical teschenite forms the Salisbury Craigs Sill on the outskirts of Edinburgh, and part of Gullane Hill on the Forth. The most distinctive of these rocks, so far as appearance in the hand-specimen is concerned, are the essexites, by reason of the large number of euhedral, porphyritic titanaugites they contain.³ An olivine-rich variety, which is correspondingly poor in feldspar and thus intermediate between the essexites and picrites, occurs at Benbeoch and elsewhere in Ayrshire, and has been termed "kylite."

The latest intrusions of the Carboniferous cycle are east-west dykes and some sills of quartz-dolerite and tholeiite, which cut, and are therefore later than, all the other intrusions. Petrographically they are unrelated to the earlier intrusions described above, and they bear no relation to the centres of Carboniferous volcanicity. They are, however, closely connected with the folding and faulting of late-Carboniferous times. They occur chiefly in the Midland Valley, and have been studied in the Kilsyth-Croy area, in the Bathgate Hills and in the neighbourhood of Stirling. Southwards they extend into northern England, and include the Hett dyke in Durham and the famous Whin Sill.⁴ In these rocks a pale brown monoclinic augite takes the place of the titanaugite of the teschenites and associated rocks: it is often accompanied by orthorhombic pyroxene (not by olivine) and sometimes by pigeonite. The feldspar is a medium labradorite. The quartz occurs in patches of interstitial micropegmatite. A typical quartz-dolerite forms the sill at Ratho, 8 miles east of Edinburgh, while an even more quartzose type occurs at Bowden Hill, Linlithgowshire. At their margins these intrusions pass into fine-grained tholeiites, of much the same composition.⁵

Summarily, the petrographic characters of the igneous rocks of Carboniferous age in the south of Scotland clearly indicate that they

¹ Tyrrell, G. W., *Trans. Geol. Soc. Glasgow*, **28**, part ii (1927-28), p. 281.

² Campbell, R., and Stenhouse, A. G., *Trans. Geol. Soc. Edin.*, **9** (1907), pp. 121-34.

³ Scott, A., *Geol. Mag.* (1915), pp. 455-61, 513-19.

⁴ Holmes, A., and Harwood, H. F., "Age and Composition of the Whin Sill," *Min. Mag.* (1928), pp. 493-542.

⁵ Walker, F., "Palaeozoic Quartz-Dolerites . . .," *Min. Mag.*, **24** (1935), p. 131.

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constitute an "oceanic" suite: the dominant basic lavas are olivine-basalts though the alkaline affinities of some are indicated by the occurrence of interstitial analcite. True andesites are unrepresented; but mugearites are interstratified with the basalts in several localities; while, as is commonly the case, trachytes, often nepheline-bearing, are among the late members of the lava succession. The intrusive phase is more obviously alkaline, with widespread teschenites and alkali-picrites carrying accessory analcite.

The regional east-west dyke-swarms which form such a striking feature of geological maps of the Midland Valley of Scotland are quite different: they represent a tholeiitic magma-type. The dykes

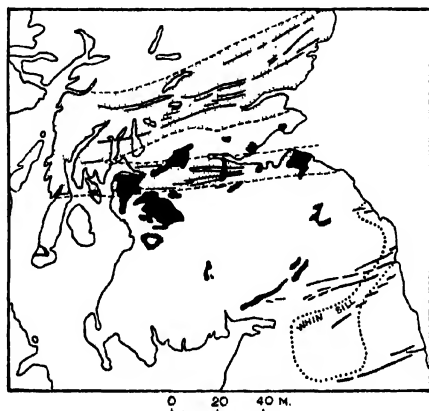


FIG. 140

Sketch-map showing Carboniferous volcanic rocks (black) and quartz-dolerite dykes in Scotland and North England. The dykes occur in the two areas dotted in Scotland; only a few of the dykes are shown. (Based on maps by Dr. F. Walker and Prof. A. Holmes.)

consist of either tholeiites (the finer grained ones), or quartz-dolerites, of the same composition but coarser grain. It is interesting to speculate on the causes of the change of magma-type, from "oceanic" to "continental," at a time when the area was being warped and fractured by the Armorican earth movements. The only certain fact is that it was a regional change and clearly indicates the advent of magma drawn from a new and different source.

(2) ENGLAND

The most extensive tracts of Carboniferous volcanic rocks in England occur in **Devon** and **Cornwall**, where the rocks form a direct continuation of the Devonian cycle, so that no further

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description is necessary. Emphasis may, however, be laid upon the fact that there is very little in common between the lavas and intrusions of this and the Scottish region: both series are essentially basic in composition; but in the case of the Scottish rocks primary differentiation had not removed the composition of the magma far from that of average basalt, although the tendency to migrate towards the "alkali pole" becomes noticeable towards the close of the cycle. On the other hand, the magma that gave rise to the spilites of the Devonian-Carboniferous cycle of South-West England was much more definitely alkaline (sodic) at the commencement of the cycle, and this character persisted throughout the cycle, with little appreciable alteration in the composition of the magma.

In **Derbyshire** igneous rocks of Lower Carboniferous age occur in the *Dibunophyllum* zone in the neighbourhood of Matlock, Miller's Dale and Tissington. The lavas include basalts of two different kinds. Sheets of normal olivine-basalt occur in multiple flows with more alkaline types.¹ In the latter the dominant feldspar is oligoclase, together with some orthoclase, while augite is usually subordinate to olivine. These lavas are therefore mugearites. They occur, as do the mugearites of the Midland Valley of Scotland, in close association with normal olivine-labradorite-basalts, and may possibly be regarded as an alkaline scum drawn off from the top part of a basalt-filled magma-basin.

The lavas are associated with tuffs and numerous sills of olivine-dolerite, and were presumably erupted from central volcanoes, since agglomerate-filled necks have been found in their vicinity. Locally the igneous rocks are termed "toadstone."

In the **Isle of Man** around Castletown the highest rocks referred to the Lower Carboniferous are porphyritic olivine-basalts, tuffs and agglomerates.

In northern **Somerset** a feeble development of basic lavas and associated intrusions is found in the neighbourhood of Weston-super-Mare. As in Derbyshire, the lavas occur high up in the Carboniferous Limestone, and are interbedded with normal marine sediments. They bear some resemblance to the spilites of Devon and Cornwall, but their state of preservation leaves much to be desired.

In the English Midlands basalts are associated with Carboniferous Limestone in the neighbourhood of Little Wenlock near the Wrekin, Shropshire, and with the Upper Carboniferous at Rowley Regis, Barrow Hill,² Pouk Hill, Kinlet, and the Clee Hills. Petrologically these rocks are closely similar, often extremely fine-grained analcite-bearing olivine-basalts, that is, analcite-basanites. Although some

¹ Sargent, H. C., *Quar. Journ. Geol. Soc.*, 73 (1918), p. 21.

² Marshall, C. E., "The Barrow Hill Intrusion, S. Staffs.," *Quar. Journ. Geol. Soc.*, 101 (1946), p. 177.

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are definitely intrusive, others are just as clearly extrusive.¹ Thus the Titterstone Clee basalt is a sill; the Rowley Regis basalt is laccolithic; the Barrow Hill mass is intrusive. On the other hand the Little Wenlock basalt, which is finely exposed in quarries at Doseley, has been converted in its upper parts into a typical red bole, due to contemporaneous subaerial weathering. Although much of the Etruria Marl in the Upper Coal Measures in the northern Midlands superficially resembles red bole, the suggestion that it was formed of weathered basaltic material has not been generally accepted,² but in certain fine-grained breccias in the Etruria Marl (the so-called espley beds) abundant shards of volcanic material may well indicate the reality of eruptions at this time.

THE ARMORICAN INTRUSIONS

Just as the formation of the "Caledonides" was accompanied by the uprise of the "Newer Granites" and the intrusions associated with them, so the birth of the Armorican or Hercynian chains coincided with the emplacement of the granite masses of South-West England and the deeply denuded mountain belt stretching through "Armorica" into Spain. Five great laccoliths or bosses and several smaller satellitic intrusions occur in Devon and Cornwall: the former comprise the granites of Dartmoor (240 square miles), Bodmin Moor (or Brown Willy) (75 square miles),³ St. Austell (33 square miles), Falmouth (or Carn Menellis) (50 square miles), and Land's End (75 square miles); while the Scilly Isles represent the highest points of a sixth large mass. The smaller intrusions include those of St. Michael's Mount, Godolphin, Carn Brea and Carn Marth, Belovely Beacon, Kit Down and Hingston Down. These apparently isolated outcrops are connected underground and there is no doubt that the intervening stretches of killas are in some cases merely roof-pendants, and that further unroofing will extend the area of granite and diminish that of the country-rock. The granites occupy a belt of country having a Caledonoid trend; but they are individually associated with Armorican axes of uplift, and were injected into the crust near a line of weakness bordering the ocean.⁴ This weak tract not only determined the positions of the Armorican intrusions, but also the location of the Devonian-Carboniferous igneous rocks, and probably also of the Permian Exeter lavas. The Armorican

¹ Pocock, R. W., "The Age of the Midland Basalts," *Quar. Journ. Geol. Soc.*, **87** (1931), p. 1; but compare Marshall, C. E., "Field Relations of the Basic Igneous Rocks associated with the Carboniferous Strata," *Quar. Journ. Geol. Soc.*, **97** (1942), p. 1.

² Robertson, T., "The Origin of the Etruria Marl," *Quar. Journ. Geol. Soc.*, **87** (1931), p. 13.

³ Ghosh, P. K., *Min. Mag.*, **21** (1927), p. 285.

⁴ Dewey, H., *Proc. Geol. Assoc.*, **36** (1925), p. 109.

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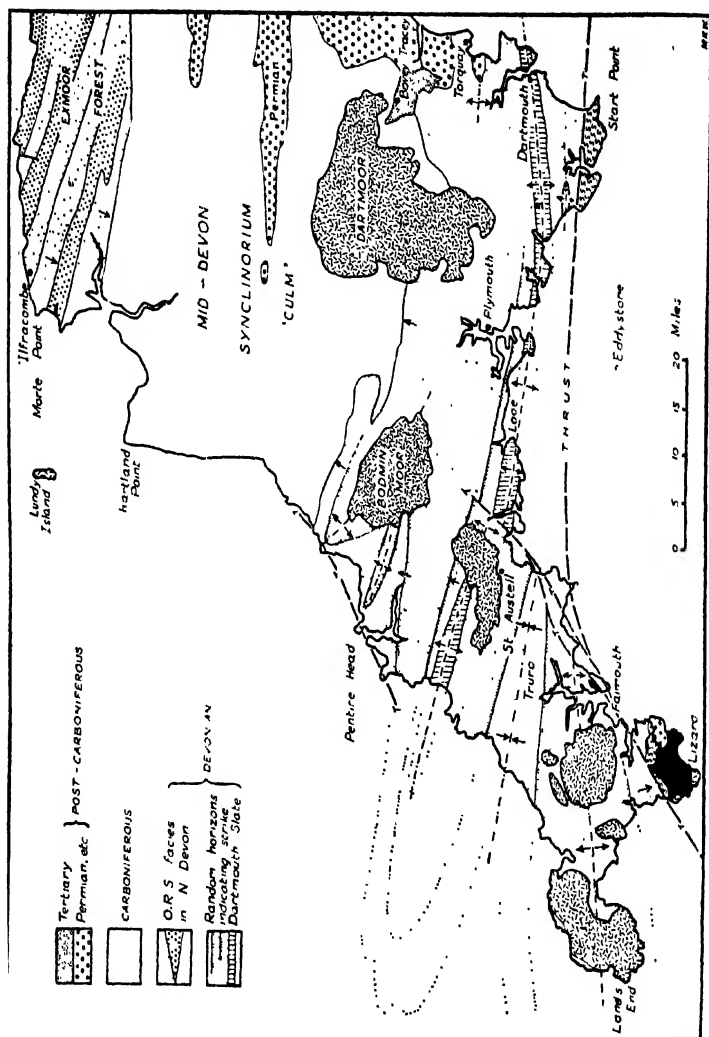


FIG. 141
Structural map of Devon and Cornwall showing the Armorican granitic complexes. (After Miss Hendricks.)

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granites of this country are important not only from a petrological, but also from an economic point of view, because their intrusion was accompanied by a period of mineral emanation, when lodes were formed containing ores of tin, copper, lead, zinc, iron, molybdenum and arsenic. To the same period belongs the intense local alteration of the granites which resulted in the very valuable deposits of china-stone and china-clay rock for which parts of Cornwall are famous. The Armorican complexes have been subjected to detailed examination, and Dr. A. Brammall's papers on the Dartmoor granite convey the results of years of intensive study.¹ Their importance in the study of petrogenesis has been referred to above.

Such evidence as had been collected with reference to the form of the intrusions suggests a laccolithic rather than a batholithic habit: the fact that the granites appear to dip gently under the altered sediments into which they were intruded implies a domed roof; while in some cases, for example Dartmoor, a stratiform arrangement of the various intrusions that make up the mass suggests a sheet-like form. There is, however, little evidence as to the nature of the floor of the intrusions, and the petrographical similarity between the several granites is so pronounced that community of origin is undoubted, while the differences may well be due in part to varying depths to which they have been denuded.

It is evident that, like the Caledonian granites, those of Armorican age are composite intrusions. The Dartmoor granite consists of three chief members which were intruded in the order of decreasing basicity.² The earliest intrusion was basic in composition, and is represented by numerous inclusions, some of large size, in the later intrusions. Next, the so-called "giant granite" was injected under a roof of killas, basic lavas, etc., and was later split into two or more sheets by a somewhat more acid "blue granite," which is a valuable building-stone. Minor intrusions in the form of narrow dykes, veins and thicker sills are common in both of the main intrusions. They are of granitic composition, but in some cases show wide variation from the main type.

The so-called "giant granite" is very coarse in texture, and rich in exceptionally large feldspar phenocrysts, which may measure 7 by 5 inches and consist of coarse microperthite. In the ground-mass, microperthitic orthoclase is associated with subhedral plagioclase ranging from albite to oligoclase. The dominant coloured mineral is biotite, with which is associated a subordinate amount of muscovite. Accessories are very variable, and in addition to those of normal occurrence in granites, such as zircon, magnetite and apatite,

¹ See particularly "The Dartmoor Granite," *Quar. Journ. Geol. Soc.*, **88** (1932), p. 171 (with Harwood, H. F.).

² Brammall, A., and Harwood, H. F., *Min. Mag.*, **20** (1923), pp. 39-53.

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include others of pneumatolytic origin, namely, tourmaline, topaz, anatase and brookite, as well as minerals resulting from assimilation of country rock by the granite. Among these are abundant almandine garnet together with cordierite, andalusite, sillimanite, corundum and spinel. The "blue granite," which was intruded beneath the "giant granite," and was thereby protected from contamination by assimilation, is practically free from these highly aluminous accessories.

The minor intrusions connected with the granites include basic as well as aplitic and pegmatitic "differentiates" of the main types. One of these, the Bittleford pegmatite on Dartmoor, contains a small amount of gold and silver of pyrogenetic origin. Of equally wide distribution are the many dykes of microgranite (some of them porphyritic), known to the miners as "elvans." Generally the aplites are more leucocratic than the main granites and more subject to pneumatolytic alteration, with tourmaline in place of biotite. The best-known aplite, which is quarried at Meldon, north of Dartmoor, is of this type, and contains a notable amount of topaz, wolfram, etc.

The dykes in Cornwall are stated to be arranged radially to the granite bosses, but locally there is a pronounced parallelism in their trend. Very few are found in the granites themselves: most occur in the synclinal tracts between the anticlines into which the granites were intruded; but there is reason to believe that the plutonic rock lies at no great distance beneath the surface in the areas where the dykes are concentrated.

The most widely distributed type of dyke-rock, well exemplified by the Prah elvan, is a strongly porphyritic variety containing numerous feldspar phenocrysts up to one inch in length, together with bipyramidal quartzes and hexagonal biotites.¹ In the elvans associated with the St. Austell granite, porphyritic muscovite is not uncommon, while topaz, although widely distributed as an accessory, is especially characteristic of the intrusions near the Land's End mass. The dykes have suffered the same pneumatolytic modifications as the parent intrusions: they are often found to have been tourmalinized and kaolinized—in some cases tourmaline is the dominant coloured constituent.

The St. Austell granite is economically the most important of the Armorican granites. Having suffered more severely from the effects of pneumatolysis than any of the others, its deposits of china-clay rock and china-stone are more extensive, although they are not absent from the other masses; while veins of greisen and schorl-rock are common. The active agents which effected the alteration of the granite rose along vertical or highly inclined joints in the granite.

The metalliferous lodes occupy a broad belt of country embracing

¹ "Land's End District," *Mem. Geol. Surv.* (1907), pp. 61-7.

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the northern half of the Land's End mass, the Redruth-Truro district north of the Carn Menellis mass, the St. Austell granite, the southern part of the Bodmin Moor intrusion, and terminates against the Dartmoor granite in the neighbourhood of Tavistock. Within this belt the dominant direction of the lodes is north-east to south-west, but south of Bodmin Moor it is north-south, and in the extreme east of the belt nearly east-west. The only important tin-mining area at the present time is that lying on the north-west side of the Carn Brea granite. From the petrological point of view, some of the most interesting of the metalliferous deposits are the wolfram-pegmatites, which are of very coarse grain, and are typically developed near Buttern Hill to the north of the Bodmin Moor granite. Here microcline crystals up to three inches in length are intergrown with wolfram and quartz, while other constituents are apatite and tourmaline. These pegmatites often pass into veins of wolfram and quartz, and finally into pure quartz-rock. A detailed consideration of the genesis of the ores is outside the scope of this book, but it is interesting to note that a zonary distribution of the several ores, dependent on the temperature of their formation, has been demonstrated.¹

The relation of the Armorican granites to the Exeter lavas is an interesting question, and it is probably identical with that connecting the Caledonian intrusions of Scotland with the Lower Old Red Sandstone lavas. Both series of lavas were extruded under sub-aerial, desert conditions; both rest upon an eroded land-surface from which they are locally separated by a small thickness of coarse arenaceous deposits; both were preceded by a period of plutonic injection.

In the Channel Islands granites are magnificently exposed in cliff sections in Jersey, Guernsey and Sark. In Jersey² the plutonic rocks include a wide range of types, red-weathering alkali-granite forming the north-western, south-western and south-eastern bastions. The north-western granite has stoped its way into earlier basic intrusions and lavas, and is consequently crowded with xenoliths in all stages of dissolution. Because of assimilation of some of this basic material the red granite, which is normally deficient in coloured minerals and of leucogranitic type, is converted into a grey porphyritic rock, with clots and schlieren of hornblende and biotite. The basic xenoliths are now of dioritic or monzonitic composition—consisting essentially of hornblende and plagioclase, with quartz, orthoclase and biotite. Normally the original structure has been destroyed; but rarely small patches are seen to have survived, and

¹ See Dewey, H., *Proc. Geol. Assoc.*, 36 (1925), pp. 107-35, and references to the previous work of Maclaren, Cronshaw and Davidson.

² Wells, A. K., and Wooldridge, S. W., *Proc. Geol. Assoc.*, 42 (1931), p. 178.

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these consist of labradorite and titanite in ophitic relationship. Therefore some of these Jersey diorites are amphibolitized (olivine-) gabbros. But in addition, dioritic rocks are widely developed in the north and south-east of the island, and include the same types of xenoliths as the granites. These are *magmatic diorites*, though they may well have been produced from the same materials as the *metasomatic diorites*, though more extensively, at some deeper-seated source. They include types with brown and green hornblendes varying in size between needle-like crystals to stout prisms several inches in length, showing a preferred orientation and particularly striking in certain pegmatitic facies. The proportion of black hornblende to white feldspar varies and some varieties closely resemble the appinitic suite of the Scottish Highlands. Arrested stoping with widespread hybridization, the development of local pegmatitic clots and aplitic veins are well displayed at Elizabeth Castle, St. Helier, and much more extensively in the extraordinary maze of reefs laid bare at low tide off the south-eastern corner of Jersey.

The plutonic rocks are cut by sills, rarely of dolerite, more commonly of acid types—granophyre and microgranite, and large numbers of dykes of basalt, dolerite and mica-lamprophyre.

In addition to the Armorican granites exposed at the surface another is believed to underlie the Alston Block in the northern Pennines.¹ The strongest evidence for its existence is a large negative anomaly proved during an electromagnetic survey of the region. The buried Alston granite is mentioned at this point, though there is no direct proof of its age: it *might* be Tertiary.

PERMIAN

Igneous rocks of Permian age are found in only two areas in Britain—in Devonshire and in Ayrshire. The latter area has already been dealt with earlier in this chapter. The lavas of the Mauchline area are associated with red sandstones usually considered to be Permian.

In **Devonshire** the Permian lavas are restricted to the neighbourhood of Exeter, and are often referred to as the "Exeter traps." They lie at, or near, the base of the "New Red Sandstone," either directly on the Culm, or separated from it by a thin stratum of red rocks. The lavas were erupted over a very irregular surface, from which rose prominent hills of Carboniferous (Culm) rocks. The main outcrops occur in the "tongues" of New Red Sandstone that extend westwards from the neighbourhood of Killerton and Tiverton. In

¹ Bott, M. H. P., and Masson-Smith, D., "The Geological Interpretation of a Gravity Survey of the Alston Block . . .," *Quar. Journ. Geol. Soc.*, 113 (1957), pp. 93-118.

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addition, small patches of lava associated with Permian conglomerates rest on Devonian slates to the south of Dartmoor near Kingsbridge, and also south-west of Plymouth. The actual centres of eruption have been located in some cases. In others the lavas are regarded as of puy type. It is possible that other vents are hidden beneath the Permian rocks, but some evidence has been adduced which indicates derivation from the high land in the neighbourhood of Dartmoor. It is possible that the Dartmoor granite itself occupies the site of a magma-basin, capped with a roof of Culm sediments, from the upper surface of which rose several volcanic cones. In the unroofing of the granite mass the cones have of course been destroyed, but it is significant that blocks of acid and intermediate lavas and strongly porphyritic microgranite occur in the Permian breccias, and decrease in size away from Dartmoor.

The systematic examination and identification of the rocks is rendered difficult by their extremely altered condition, the alteration dating from the Permian period. The lavas were the product of sub-aerial eruptions, and were extruded under desert conditions. Hence they suffered rapid disintegration, and their iron-bearing constituents were quickly oxidized, which gives the rocks a characteristic red colour. The upper portions of the lavas are often much fractured (block lavas), and the crevices filled with sand. Many of the flows are extremely vesicular, and in their general appearance are strongly reminiscent of the Old Red Sandstone lavas of the Pentland Hills, etc., and the Permian lavas of Ayrshire.

The Exeter lavas have been investigated by Tidmarsh,¹ who divides them into ten types grouped into three series. Each of the types exhibits several different facies. Among the lavas and associated intrusions are some to which the names of normal rock-types may be applied, such as minette, rhyolite, "quartz-porphry." Most of the rocks, however, are of quite exceptional mineral and chemical composition. Among the features of special interest are: the occurrence of quartz-xenocryst-basalts; the lamprophyric facies of the lavas; and the occurrence at Loxbeare near Tiverton of a biotite-olivine-leuconite which is evidently an under-saturated minette. The leucites are small but typical, and show the regular groupings of minute inclusions so characteristic of the small leucites in chilled lavas. Apparently rapid chilling of the magma at a time when mica was in the process of formation gave rise to olivine plus leucite instead of more mica.

The problem of the origin of these Exeter lavas is the problem of the genesis of lamprophyres. Tidmarsh's suggested solution involves interaction between the "depth residuum" of the Dartmoor granite

¹ Tidmarsh, W. G., "The Permian Lavas of Devon," *Quar. Journ. Geol. Soc.*, 28 (1932), p. 712.

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on the one hand, and a late acid residuum on the other. Both consisted partly of crystals and partly of magma, therefore both sets of crystals would be unstable in the mixed liquid and would suffer corrosion in the manner described.

Although the range of composition of the Exeter lavas is small,

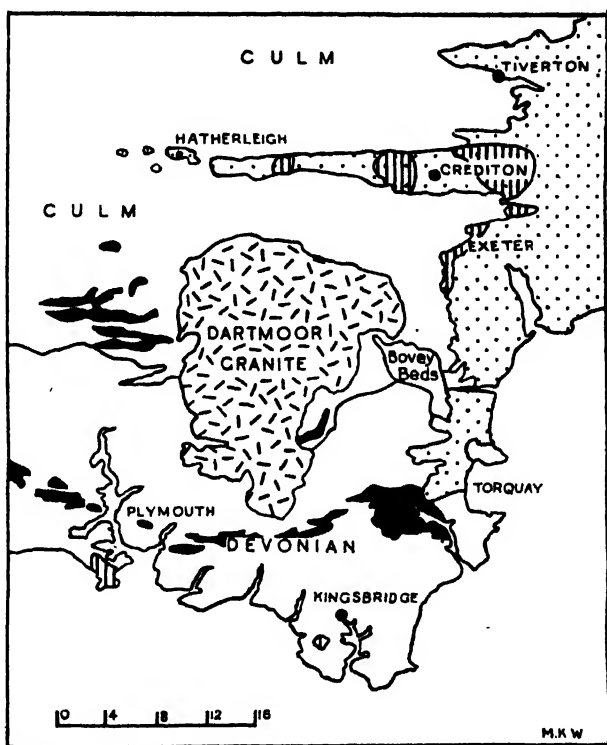


FIG. 142

Sketch-map showing distribution of the Exeter lavas and Devonian-Carboniferous volcanic rocks in Devonshire. (After W. G. Tidmarsh.) Permian lavas, spots; areas containing Permian lavas, ruled; Devonian and Carboniferous volcanic rocks, solid black.

it is significant that the Permian breccias contain abundant pebbles of strongly porphyritic microgranite rich in large phenocrysts of quartz and feldspar, also rhyolites and spherulitic acid lavas and andesites.

These Permian igneous rocks of South-West England form a small portion of a large petrographic province extending into Germany.

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On the Continent, however, the amount of igneous material is much larger and more varied in composition.

TERTIARY

Introduction

Throughout the whole of the Mesozoic Era there were no volcanic outbursts in the British area; but during the Tertiary Era igneous activity occurred over an enormous area known as the **Brito-Icelandic** or **Thulean** Province, which embraces North-West Britain, and extends northwards into Iceland, Spitzbergen, the Faeroe Islands and Jan Mayen, and westwards to Greenland. In Britain volcanic activity was confined to the early Tertiary, ending possibly 40 million years ago; but in Iceland activity has persisted until the present time.

The distribution of the Tertiary volcanic rocks in Scotland and Northern Ireland is shown in Fig. 143. Extensive tracts of plateau-building and flood-lavas are preserved in Antrim and in the Inner Hebrides, particularly in the northern part of Skye, and on the island of Mull and the adjacent mainland. In addition centres of localized igneous activity occur in Skye, Mull, the island of Rum (or Rhum, one of the Small Isles of Inverness-shire), on the Ardnamurchan peninsula, in central and northern Arran, and in three localities in Northern Ireland, namely the Mourne Mountains, Slieve Gullion and Carlingford. Since the end of the period of volcanic activity erosion has removed much of what must originally have been very extensive lava fields and has exposed the intrusive rocks of the volcanic centres listed above. Some, and perhaps all, of the latter represent the roots of volcanoes of central type, many of which had a long and complex history of activity. Because of fortunate circumstances of erosion and exposure, and the fact that it has been studied intensively by a number of distinguished geologists, it is probable that the structure and history of the Mull complex are better known than those of extinct volcanoes from any other part of the world.

The Tertiary igneous rocks have not suffered subsequent folding or metamorphism, and hence they are usually in an excellent state of preservation. The lavas can be examined in hundreds of miles of coastal sections, while the rocks of the various intrusion centres are generally well exposed in areas of considerable relief. It is not surprising, therefore, that the phenomena of Tertiary igneous activity have attracted much detailed research. The reader is advised to study the valuable summary provided by Dr. Richey¹, and in addition reference should be made to the classic Survey Memoirs

¹ Richey, J. E., "Scotland: The Tertiary Volcanic Districts," *British Regional Geology*, 2nd Edn., 1948.

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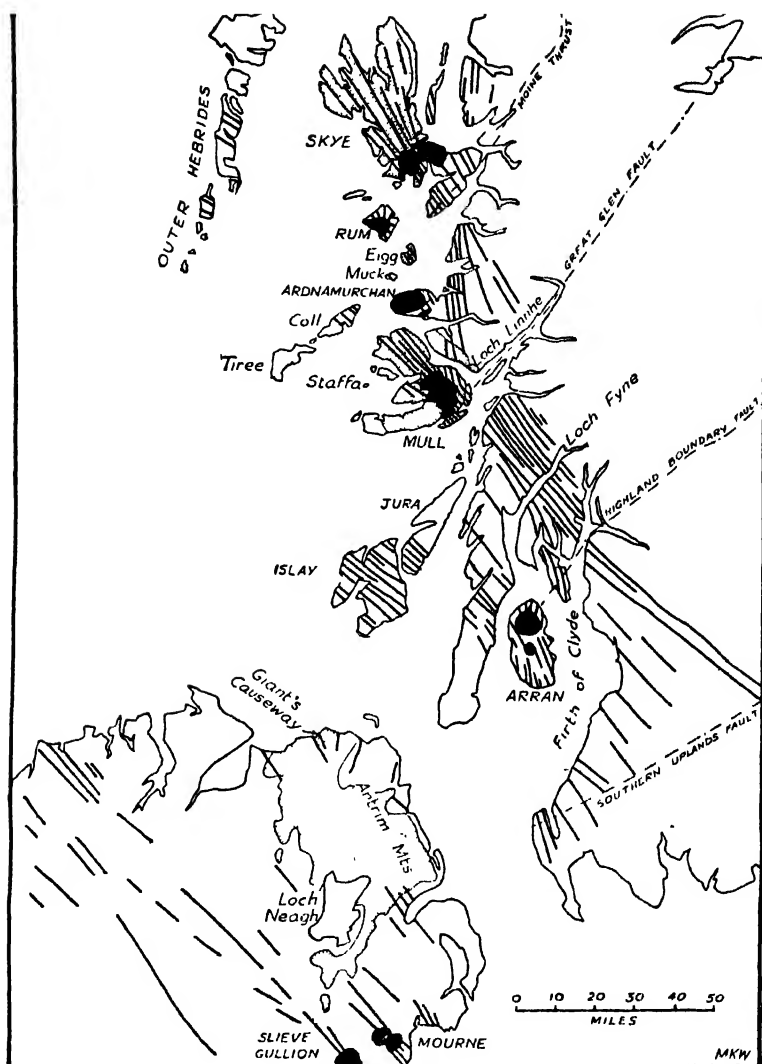


FIG. 143

Map of the Tertiary igneous rocks of Scotland and northern Ireland.

(Based on maps by the officers of the Geological Survey.)

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and accompanying 1-inch maps on Skye, Mull, Ardnamurchan and Arran.¹

The Tertiary igneous cycle consists of the usual three phases: the eruptive (extrusive) phase, the phase of major intrusions, and that of the minor intrusions—the dyke phase.

The Eruptive Phase: the Lavas

Lavas of Tertiary age in Britain are overwhelmingly basaltic; those of "intermediate" composition are practically unrepresented; but rhyolitic lavas and plugs occur rarely. It is a little difficult to deal with the basalts adequately on account of some confusion which has arisen in the use of the terms 'type,' 'series,' 'groups,' 'modes of occurrence' and 'magma-types.' From the petrographic angle it appears that the basalts fall into two main types corresponding with those recognized by Kennedy and Anderson, and discussed above under "origin of basalt." These are (1) over-saturated, olivine-poor, fine-grained basalts of tholeiite type, containing interstitial glass; and (2) olivine-basalts, under-saturated as regards mafic constituents, containing some 20 per cent of olivine, though this figure may rise to up to 50 per cent in certain melabasalts of picrite-basalt type. Augite-rich melabasalts (ankaramites) also occur, but are much more important in the outlying parts of the region, for example Jan Mayen, where they dominate the basal group. Among the olivine-basalts aphyric and strongly porphyritic varieties are distinguished. Less common types of the latter contain very prominent phenocrystic plagioclases up to 3 inches in length and termed "Big Feldspar Basalts" by the Survey. They belong in general to the Markle type defined above. Other paler coloured basalts are rich in microphenocrysts of plagioclase, while mugearites occur occasionally as flows, as well as minor intrusions (in Skye, the type locality).

The lavas were erupted under subaerial conditions and on account of the general absence of pyroclastic rocks it is inferred that there was little explosive volcanic activity, though several vents filled with agglomerate or even merely brecciated Chalk have been discovered in Antrim and evidently represent volcanoes of phreatic or steam-blast type.

¹ *Memoirs of the Geological Survey* relating to Tertiary igneous activity in Scotland include:

Harker, A., "The Tertiary Igneous Rocks of Skye" (1904);

Bailey, E. B., Clough, T. C., Wright, W. B., Richey, J. E. and Wilson, G. V., "The Tertiary and Post-Tertiary Geology of Mull, Loch Aline and Oban" (1924);

Tyrrell, G. W., "The Geology of Arran" (1928);

Richey, J. E., and Thomas, H. H., "The Geology of Ardnamurchan, North-West Mull, and Coll" (1930).

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In northern Antrim the lava succession¹ is interrupted at two levels by well-developed laterite horizons representing periods of quiescence during which the exposed basalts were deeply weathered. Elsewhere in North-East Ireland only one Interbasaltic Bed occurs. In addition individual basalts may have weathered reddened tops which in some cases are red lateritic clays or boles. Terrestrial conditions are further evidenced by very occasional plant-beds containing the impressions of leaves; by the preservation of trunks of trees overwhelmed by, and engulfed in basalt; and by very thin lignite or coal seams, including one associated with a few feet of Tertiary mudstone underlying the lavas of Mull, Morvern and Ardnamurchan. The leaf-beds, of which that occurring at Ardtun in South-West Mull is the best known, provide the only direct evidence of the age of the lavas, and of the climatic conditions at the time. Actually the plant-remains are not sufficiently distinctive to fix the age within the Tertiary Epoch precisely, but the palaeobotanists concerned are agreed that they probably represent an early Eocene flora. A general indication of Tertiary age is shown by the stratigraphical relationships: in Antrim the lavas rest directly on an eroded surface of Upper Chalk; while in the Hebrides they overlap on to other Mesozoic and older strata.

With regard to general features, the lavas vary in thickness from a few feet in the case of certain ropy (pahoehoe) lavas to 150 feet, the average being perhaps 40 feet. Although the central parts of the flows are massive (and indistinguishable in this respect from basaltic minor intrusions) the upper parts are often slaggy and extremely vesicular, so that they weather readily, giving rise to a terraced type of scenery. Mapping is greatly assisted by this terracing, and it has proved possible to trace individual flows and to prove that they maintain their thicknesses in many instances for long distances.

Some of the Tertiary basalts are world famous for the magnificent columnar structure which they display, particularly at the Giant's Causeway in Co. Antrim,² and at Fingal's Cave in the Hebridean island of Staffa. On a point of mineralogical detail it may be noted that at certain localities (including the Giant's Causeway) the lavas are strongly vesicular and provide well-crystallized specimens of many varieties of zeolites and other minerals for collectors.

Some difficulty has been experienced by surveyors in distinguishing between the more massive lavas and sills of the same

¹ Patterson, E. M., "The Tertiary lava succession in the northern part of the Antrim Plateau," *Proc. Roy. Irish Acad.*, 57 (1955), p. 79. A useful summary of the Tertiary igneous rocks in North-East Ireland may be read in *Proc. Geol. Assoc.*, 71 (1960), p. 441.

² Tomkeieff, S. I., "The basaltic lavas of the Giant's Causeway District of Northern Ireland," *Bull. Volc. Naples, Série ii, Tome 6* (1940), p. 89.

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petrographic type. This was notably the case in Skye where a group of basic "sills" were proved later to be extrusive. Genuine basic sills do occur, however, sometimes underneath the volcanic pile, sometimes intruded into it. Noteworthy examples are the 300-foot sill (which is of sufficiently coarse grain to be classed as gabbro), forming Fair Head in North-East Ireland, and the 100-foot Portrush Sill.

An additional complication is the occasional occurrence of composite flows with sharply defined and sometimes pseudo-intrusive contacts between the lower and upper parts of the lava. Such basalts have been described from Skye¹ and Northern Ireland.²

The Interbasaltic Horizons automatically divide the lavas of northern Antrim into three series of which the Upper and Lower Series are petrographically alike, consisting of olivine-basalts; but the Middle Series consists of tholeiites. In parts of Scotland the lavas are divided on a different basis into (1) the regional basalts and (2) the products of local volcanoes of "central" types, of which the most important is the Mull Volcano. The regional basalts of the whole province are comparable in status with the flood-basalts of the Parana Basin or the Deccan. In the topographical sense they are plateau-building and constitute the Plateau Group of the authors of the Survey Memoirs to which attention was drawn above. The lavas erupted by the Mull Volcano or any other volcano of the same type constitute localized Central Groups. Only relics of the original basaltic plateau have survived; but even so they occupy approximately 2,000 square miles, of which some 1,500 square miles occur in North-East Ireland where they are magnificently exposed on the Antrim coast. The thicknesses which have survived vary widely from a maximum of 6,000 feet in Mull (equally divided between the regional and the local groups), 2,000 feet in Skye, but only 300 feet in Ardnamurchan. The lavas of the Mull Central Group were erupted from a volcano which changed in character during its long history. At one important period it was a great caldera the dissected remains of which occupy the south-eastern parts of the island. Within the caldera some of the lavas are unique among the British Tertiary basalts in exhibiting pillow structure which proves that they were erupted into a crater lake at that time occupying the caldera depression.

During its early active life the Mull Volcano must have been in many ways similar to the existing volcano of Hawaii: a broad shield built up mainly of basaltic lavas and with a summit caldera, from the rim of which the main effusions of lava took place. There must

¹ Kennedy, W. Q., "On composite lava flows," *Geol. Mag.*, 68 (1931), p. 166.

² Walker, G. P. L., "Some observations on the Antrim basalts and associated dolerite intrusions," *Proc. Geol. Assoc.*, 70 (1959), p. 179.

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have been many other volcanoes scattered over the lava plateau in early Tertiary times, the sites of some of the more deeply dissected being now represented by the intrusion complexes which form the subject of the next part of this account.

It seems probable that in the reservoir underlying and feeding the Mull Volcano, basalt magma was locally differentiated to a greater extent than was normally the case for the plateau region as a whole. Various genetic relationships have been suggested between the "plateau" and "central" types of magma involving, amongst other factors, fractional crystallization with the formation of some magma fractions depleted in crystals and others with crystals added.

Similar local differentiation has produced lavas of Intermediate and Acid composition near to the intrusion centre of the Cuillin Hills of Skye. This centre must have been the site of a volcano similar to that of Mull at the time of the lava eruptions, and gave rise to up to 2,000 feet of trachytes and overlying rhyolitic tuff, agglomerate and lava.

In the Central Complex of Arran, the preservation of blocks of Mesozoic sedimentary rocks (which are otherwise unrepresented in the island) suggests the former presence of a collapsed caldera. Some of the blocks may have been down-faulted about 3,000 feet. Within the caldera, volcanic cones were built up of andesite, dacite and rhyolite and their pyroclastic equivalents.¹

There is no doubt that lavas of the so-called Central Group were erupted from the Mull Central Volcano, and the same mode of origin is probable for other lavas of restricted distribution around other plutonic centres (*e.g.*, the Cuillins of Skye); but much more uncertainty surrounds the source and manner of eruption of the more extensive flows of the so-called Plateau Group. Sir A. Geikie long ago suggested that the latter might have been erupted from fissures; but until recently no proof of a direct connection between feeding dyke and a lava flow had been forthcoming. Furthermore the members of the Tertiary dyke swarms which would provide the obvious feeding channels for the lavas are in the great majority of cases younger than the lavas which have survived denudation. On general grounds, however, it cannot be doubted that many of these dykes must have reached ground level and therefore acted as feeders during fissure eruptions. It is inconceivable that parallel, presumably tensional, fissures could be restricted to the *lower* levels of the basaltic plateau. The chances of seeing a direct link between any one of these hundreds of dykes and a particular basaltic lava must be extremely slight; yet it has proved possible to demonstrate that fissure eruptions did actually occur in

¹ King, B. C., "The Ard Bheinn area of the Central Igneous Complex of Arran," *Quar. Journ. Geol. Soc.*, **110** (1954), pp. 323-55.

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North-East Ireland when the basaltic plateau was in process of formation.¹ About thirty dolerite plugs occur within the area of the Antrim Plateau, and these are thought to have formed by enlargement at points along fissures through which basaltic magma was passing. Fissure eruptions have occurred within historic times in Iceland, from the Laki Fissure for example.

The Phase of Major Intrusions

The chief interest of the British Tertiary Igneous Cycle lies undoubtedly in the unique variety of ring-dykes and cone-sheets which collectively make up the central complexes. The ring-dykes and all the larger bodies such as bosses and stocks associated with them are described for convenience as "major" intrusions, although in fact they vary greatly in size and in the coarseness of texture of their component rocks. All these intrusions constitute the basal structures of volcanoes of central type exposed by the removal of the cover of ejected material—surface lavas and tuffs—under which they consolidated. Evidently the rocks of the intrusive phase rose into the base of the volcanic piles which had accumulated during the eruptive phase. It is impossible to estimate the thickness of the cover; but it should be remembered that the *total* thickness of lavas locally reached 6,000 feet. Frequently the intrusive rocks are now exposed at the same levels as the lavas which must have flanked the original volcanoes. None of the intrusions is "plutonic" in the sense of being of really deep-seated character, though most of the rocks involved fall in the coarse-grain category.

In the main the rocks forming the major intrusions are either thoroughly basic or thoroughly acid—intermediate types are virtually absent. The close association of contrasted acid and basic rocks is displayed in various ways in each of the complexes. In Skye the largest of all the basic masses—comprising the "banded gabbros" of the Cuillin Hills—forms sombre and rugged topography which contrasts strongly with the adjacent Red Hills formed of granophyres that have given rise to scree-covered, rounded mountains.

In Mull the emplacement of gabbro and granophyre tended to alternate in time, to form a complex series of mainly arcuate intrusions which encircle the principal caldera and to a lesser extent enclose a later intrusion-centre located to the north-west of the caldera. Ardnamurchan witnessed a similar alternation of gabbroic and granophyric intrusions. In this case the intrusions form concentric arcs or rings about three well-defined centres (Fig. 144)

¹ Patterson, E. M., "Evidence of fissure eruption in the Tertiary lava plateau of North-East Ireland," *Geol. Mag.*, 87 (1950), pp. 45-52. See also Walker, G. P. L., *op. cit. supra*.

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which shifted with time along an E.-W. axis. Other aspects of the close association of acid and basic rocks are referred to below.

Although a sequence of intrusion from basic to acid rocks is common in many igneous complexes it was obviously not adhered to in some of the British Tertiary centres. Only Skye and Rhum conform to the general rule: the Skye gabbros and associated

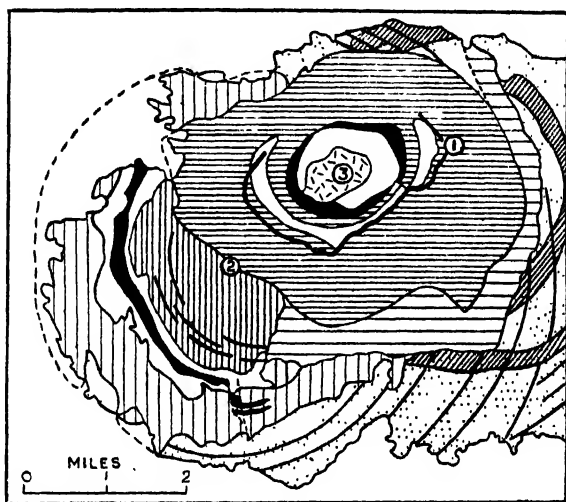


FIG. 144

Simplified sketch-map of central intrusion complex, Ardnamurchan, Scotland. Ring-dykes of Centre 1, oblique ruling; of Centre 2, vertical ruling, and black and white; of Centre 3, horizontal ruling, and black and white. Tonalite of Centre 3, small dashes. Sediments, lavas, etc., spots. A few cone-sheets round Centres 1 and 2 are shown. (Based on maps in *Ardnamurchan Memoir*, 1930.)

ultrabasic rocks preceded the granophyres; and in Rhum the sequence of intrusion was (1) anorthosites, bytownite-troctolites (allivalites) and "peridotites," (2) gabbro, (3) granophyre.

In view of the overwhelming abundance of tholeiites and olivine-basalts among the lavas which initiated the volcanic phase of activity it is to be expected that various types of chemically similar gabbros should be dominant in the major intrusions. Rather surprisingly, granites and granophyres are even more voluminous in some centres. Thus the Mourne Mountains¹ complex in Northern Ireland (Fig. 146) is composed entirely of granite forming five successive intrusions. Most of the northern half of Arran is occupied

¹ Richey, J. E., "Structural Relations of the Mourne Granites," *Quar. Journ. Geol. Soc.*, **83** (1928), p. 658; but cf. Emeleus, C. H., in "Geology of North-East Ireland," *Proc. Geol. Assoc.*, **71** (1960), p. 446.

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by a complex of granites of both coarse- and fine-grained types. In Skye the area covered by granophyres almost equals that of gabbros; but in Rhum, Mull and Ardnamurchan, basic and ultrabasic rocks are dominant.

Since it is impossible to describe the intrusion centres systematically in the space available, certain aspects of interest are selected in relation to intrusions which cover the full composition range from ultrabasic to acid.

The largest of the ultrabasic complexes, in Rhum (Fig. 148), provides a convenient starting point for this survey. The rocks are composed dominantly of anorthite-rich plagioclase (An_{85-90}) and Mg-rich olivine in varying proportions, with chromite a notable accessory. The most conspicuous and interesting feature of this complex is the occurrence of well-defined layered units of the order of a hundred feet in thickness. Within each such unit the rock-type varies from "peridotite"¹ at the base to bytownite-troctolite ("allivalite" of Harker) above, with the development of some bands of pure anorthosite at the top. Because of the contrasts in colour and weathering properties of the alternating olivine-rich and plagioclase-rich layers, these form a very striking feature of the twin mountains of Allival and Askival which dominate the centre of the island. Harker,² in his original description of the Complex, believed that the layers were separately and successively intruded; but this has been shown to be inconsistent with the contact relations between the layers.³ The complex has now been re-described in great detail.⁴ Each layered unit resulted from the bottom accumulation of olivine crystals that settled out of the magma body as a result of their early crystallization and high density. The factors controlling the rhythmic precipitation of crystals are not well understood. It is believed, however, that the chromite, magnesian olivine and calcic plagioclase represent concentrates of the earliest precipitated minerals formed at the highest temperatures in a body of basaltic magma which may have been periodically replenished from below and also acted as a feeder for surface eruptions.

¹ It should be noted that the nomenclature of the bytownite-olivine rocks adopted by Harker and Brown does not accord with that used in this book. Brown restricts "allivalite" to varieties with more than 50 per cent plagioclase, while those with excess olivine he terms "peridotites." According to our terminology, all of these rocks would be varieties of bytownite-troctolite (either normal leuco- or mela-troctolites as the case might be): exceptions would be olivine-free layers of anorthosite and plagioclase-free layers of peridotite.

² "The Geology of the Small Isles of Inverness-shire," *Mem. Geol. Surv.* (1908), pp. 68-92.

³ Tomkeieff, S. I., "On the petrology of the ultrabasic and basic rocks of the Isle of Rhum," *Min. Mag.*, 27 (1945), p. 127.

⁴ Brown, G. M., "The layered ultrabasic rocks of Rhum, Inner Hebrides," *Phil. Trans. Roy. Soc. Lond., Ser. B*, 240 (1956), pp. 1-53.

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Olivine-gabbros and "eucrites" (of the Survey) are widely distributed and form some of the biggest ring intrusions in Skye, Mull and Ardnamurchan. It is appropriate to comment first on the gabbro forming the Cuillin Hills in Skye, since this is associated with an ultrabasic (peridotitic) facies, and the gabbros are conspicuously layered though not on such an impressive scale as Rhum.

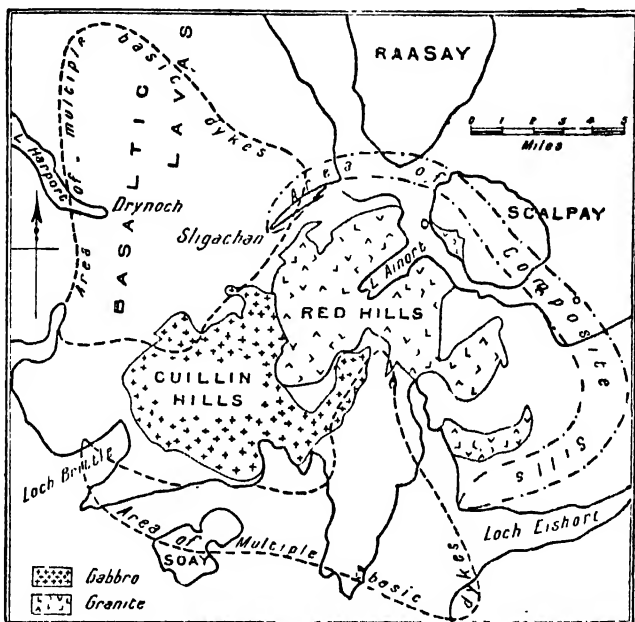


FIG. 145

Map of a portion of the Isle of Skye showing the distribution of the gabbro and granite, and of the multiple basic dykes and composite sills.

(After A. Harker.)

Layering is a feature displayed by several of the major gabbroic intrusions in Mull, Ardnamurchan, Carlingford and elsewhere. Where it is gently dipping and shows evidence of grading it can be ascribed largely to gravity stratification; but where the banding is steeply dipping, as in parts of the Hypersthene-Gabbro of Ardnamurchan, it must be due largely to differential flow movements in the crystallizing magma.¹

It is worth noting that even if the layering in these rocks is due either largely or in part to bottom accumulation of crystals, the

¹ Wells, M. K., "The structure and petrology of the Hypersthene-Gabbro intrusion, Ardnamurchan, Argyllshire," *Quar. Journ. Geol. Soc.*, **109** (1953), p. 367.

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thicknesses of the layered series exposed are insufficient to show appreciable cryptic layering.

Some of the "eucrites" and olivine-gabbros are closely associated with quartz-gabbros, which may form either separate contiguous ring-intrusions (examples occur in both Centre 1 and Centre 2, Ardnamurchan); or the quartz-gabbro may be only a marginal facies grading into the main olivine-bearing gabbros (*e.g.*, the Hypersthene-Gabbro of Ardnamurchan). This is a feature that has not been satisfactorily explained.

Perhaps the most impressive of the basic ring-intrusions is the Great Eucrite of Ardnamurchan. Due to the coarse texture and massive nature of the rock it has withstood erosion to form a most striking ring of hills encircling Centre 3 (Fig. 144). The considerable width of the outcrop between the concentric inner and outer margins is difficult to account for by the mechanism of ring-faulting and central subsidence outlined on p. 146. It is quite possible that the Great Eucrite and certain other gabbroic ring intrusions possessing wide outcrops are in fact not ring-dykes in the accepted sense. It has been suggested, for example, that the Hypersthene-Gabbro of Centre 2, Ardnamurchan, has the shape of an inverted cone or funnel,¹ and that the internal banding, which is definitely of this form, may offer a guide to the external shape of the intrusion. Rather similar conclusions have been reached in regard to the Ben Buie gabbro in Mull.²

Of slightly more acid composition than the rock-types mentioned above are quartz-dolerites which form a number of interesting intrusions particularly in Mull and Ardnamurchan. These rocks are closely related in composition to many of the cone-sheets; they commonly carry phenocrysts of plagioclase and possess a mesostasis of micrographically intergrown quartz and alkali-felspar. The latter weathers out as mossy-textured white patches on the rock surface. One of the most remarkable of the quartz-dolerite intrusions is the Glen More ring-dyke in Mull, which in crossing a region of about 1,500 feet relief shows gradation in a vertical sense from quartz-dolerite (sp. gravity 3.06) at the lower levels to microgranite (sp. gravity 2.50) above.³ This has been explained as a straightforward case of gravity differentiation with the squeezing out of a residual acid magma fraction (corresponding to the mesostasis of the quartz-dolerite), and injection of the latter into the upper parts of the arcuate fracture.

Several of the quartz-dolerite intrusions are extensively veined by

¹ Wells, M. K., *op. cit.*, *supra*

² Lobjoit, W. M., "On the form and mode of emplacement of the Ben Buie intrusion, Island of Mull, Argyllshire," *Geol. Mag.*, **96** (1959), p. 38.

³ "Tertiary and Post-Tertiary Rocks of Mull," *Mem. Geol. Surv.* (1925), p. 306.

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granophyre, giving rise to the phenomenon of net-veining. This is a form of composite intrusion and demonstrates very clearly the extremely intimate association of acid and basic magmas which is so characteristic of the Tertiary province. There can be no doubt in many cases that the dolerite and granophyre magmas were intruded together, as shown by the way they share the same intrusion cavities. This is clearly demonstrated in the extensively veined Quartz-Dolerite ring-dyke of Centre 2, Ardnamurchan¹ which has a number of sills projecting from its outer margin. The vein complexes are scattered throughout all parts of this complex intrusion. Despite their simultaneous injection, the acid and basic magma fractions have only occasionally mixed to give gradational intermediate rock-types. Generally the granophyre veins cut and brecciate the dolerite quite cleanly, due presumably to the lower temperature of crystallization of the granophyre compared with the dolerite. It is interesting to note that heat has locally been transferred from the basic to the acid magma so that at the granophyre/dolerite contacts the latter is sometimes chilled.² In the Slieve Gullion complex granophyre occurs as vertical pipes in dolerite. Again the latter is chilled against the former.³

Major intrusions of intermediate composition are extremely rare in the Tertiary province. Hybrids of very heterogeneous character, collectively termed diorites by Tyrrell, have been formed by the acidification of basic rocks in the Central Complex of Arran, and to a limited extent such rocks occur also in the other centres, notably in Skye, as exemplified by Harker's "marscoite." Almost the only well-defined intrusions in this range of composition, however, occur as bosses at the focus of Centre 3 in Ardnamurchan. These are of tonalitic and quartz-monzonitic composition and show petrographic features indicating a hybrid origin.

The major intrusions of granite have been listed above. Although some possess typical granitic texture (e.g., the coarse member of the North Arran (Goatfell) intrusion) the majority are granophyric to some extent. Small drusy cavities lined with well-formed crystals of smoky quartz, alkali-feldspar, mica and occasionally rarer minerals such as beryl, are characteristic, particularly of the Mourne Mountains and Lundy granites. As explained on p. 216, the granophyric texture is widely accepted as a feature resulting from simultaneous crystallization of quartz and alkali-feldspar from a melt. A high-temperature magmatic origin has been confirmed for one of the

¹ Wells, M. K., "The structure of the granophyric quartz-dolerite intrusion of Centre 2, Ardnamurchan, and the problem of net-veining," *Geol. Mag.*, **91** (1954), p. 293.

² Personal communication from R. R. Skelhorn.

³ Elwell, R. W. D., "Granophyre and hybrid pipes in a dolerite layer of Slieve Gullion," *Journ. Geol.*, **66** (1958), p. 57.

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Skye granites¹ from the evidence of the quartz and feldspar, the latter occurring as the high temperature form of alkali-feldspar in the chilled marginal facies and changing to perthite nearer the centre of the intrusion. However, it appears that rocks possessing either genuine granophyric (or at least very similar) intergrowth-textures, may have been developed locally by metamorphism and metasomatism. Doris Reynolds² has described an interesting example in which the Caledonian granodiorite of Newry has been locally transformed into a granophyric rock adjoining the acid ring-dyke of the Tertiary Slieve Gullion Complex (Fig. 146). The change from one rock-type to the other is gradational and involves corrosion of the original minerals and their replacement by a quartz, alkali-feldspar intergrowth.

Some of the Tertiary acid ring-dykes were affected by volcanic gases which caused internal brecciation and led to the development of intrusive rocks of agglomeratic appearance.³ The nature and modes of origin of such intrusive tufts were discussed in the chapter on Pyroclastic Rocks.

It is reasonable to assume that the surface expression of ring-dykes of this kind would be the development of agglomerate-filled vents. A good example is provided by an arcuate series of such vents which partially encircle Centre 1 in Ardnamurchan.

The acid intrusions vary greatly in their form and mechanism of emplacement. In the case of a ring-dyke such as the Loch Bà Felsite of Mull (Fig. 147), the mode of intrusion is reasonably certain. This is one of the few cases where subsidence of a central block bounded by a ring fault can be shown to have occurred, the displacement amounting to about 3,000 feet. Intrusion took place after the faulting. It is possible that cauldron subsidence may have occurred to produce some of the larger granitic and granophyric masses such as those of the Mourne Mountains, and the Red Hills. Unfortunately there is very little evidence other than the fact that many of the marginal contacts are steep. There is no justification for a view that was held for many years, that the granites of the Red Hills are laccoliths. In Mull, room was created for intrusion of some of the granophyres by lateral compression, pushing aside the country-rocks to form a most remarkable series of parallel anticlines and synclines which are concentrically arranged round the intrusions encircling the Mull Caldera. An equally convincing

¹ Tuttle, O. F., and Keith, M. L., "The granite problem: evidence from the quartz and feldspar of a Tertiary granite," *Geol. Mag.*, 91 (1954), pp. 61-72.

² Reynolds, D. L., "The transformation of Caledonian granodiorite to Tertiary granophyre on Slieve Gullion, Co. Armagh, Northern Ireland," *Int. Geol. Cong.*, 18th session, Great Britain (1948), Part 3, p. 20.

³ Richey, J. E., "Association of explosive brecciation and plutonic intrusion in the British Tertiary igneous province," *Bull. Volcanologique, Napoli*, Série 2, tome 6, pp. 157-75.

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FIG. 146

Map of part of north-eastern Ireland showing the igneous complexes of the Mourne Mountains, Carlingford and Slieve Gullion.

Newry granodiorite (Caledonian), large dashes: Mourne Mountains granite, small dashes: Tertiary gabbros, stippled: granophyres, dot-dash ornament: the acid ring-dyke of Slieve Gullion, black. One cone sheet is shown round the Mourne granites, which are numbered in order of intrusion. (After Richey, Reynolds, Tomkeieff and Marshall.)

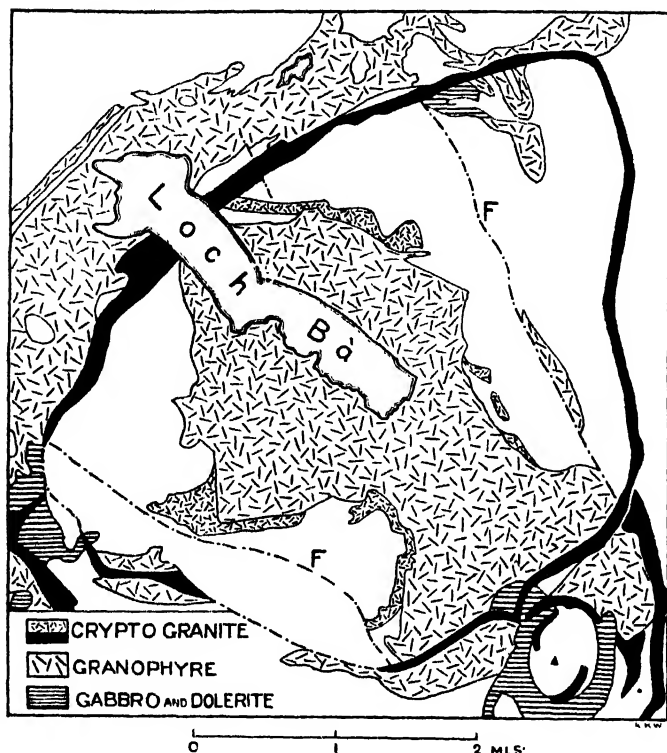


FIG. 147

A Tertiary ring-dyke and associated intrusions, Loch Bà, Mull. (After E. B. Bailey and others, in "Tertiary and Post-Tertiary Geology of Mull," *Mem. Geol. Surv.*, 1924.)

example of forceful intrusion is provided by the granites of North Arran. Uprise of the granite has forced the Dalradian and O.R.S. strata into vertical positions all round the granite so that the strike is now concentric with the margin of the latter. The mechanism in this case must have been closely analogous to emplacement of a salt dome.

The Minor Intrusions: Sills, Dykes and Conc-Sheets

The Tertiary minor intrusions, like the lavas, may conveniently be divided into two groups, (1) regional, and (2) local.

Intrusion on a regional scale took the form of swarms of basic dykes having a general north-west to south-east direction.

In connection with the local centres, the rocks of the minor intrusions were much more variable in form and in composition. Dyke-rocks are found closely allied in composition to each of the plutonic

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types noted above: there are three chief groups corresponding with the acid, basic and ultrabasic major intrusions, and in Skye they were injected in the order named.

The earliest of the post-granite minor intrusions in Skye were composite sills consisting of an earlier basic member (basalt), followed by a later acid member (graphic microgranite) which usually split the former down the centre. They are restricted to an arcuate belt lying north-east of the granite of the Red Hills (Fig. 145).

The local minor intrusions of acid composition are clearly related to the great granite intrusions, but are slightly more acid than the latter. In texture they vary considerably. Some are typical porphyritic microgranites with large phenocrysts of quartz and feldspar set in a groundmass which varies from microcrystalline to cryptocrystalline. Some of the visible offshoots from the granites are indistinguishable from rhyolites. In a parallel series the quartz and feldspar of the groundmass are intergrown, yielding micrographic, cryptographic and beautifully spherulitic rocks. Rarely these rocks become less acid, and by failure of the quartz pass into microsyenites, which in Skye and Arran closely approach trachytes in general appearance. These are the "bostonites" and "orthophyres" of Harker.

In areas near to the granites, intrusions of a late date include tholeiites and pitchstones, often associated together in composite intrusions, though both do occur alone. The pitchstones in particular are interesting on account of their rarity. First recognized by Jameson about 1700, they were more fully described by Judd, but new features, such as the occurrence of fayalite, are still being brought to light by careful examination. The best known are the pitchstones of Arran and of Eigg. In the former island they occur as sills, but more commonly as composite dykes, such as those found on the foreshore at Tormore, associated with strongly porphyritic microgranites and tholeiites.

Local minor intrusions of basic composition are in like manner related to the great gabbro and eucrite intrusions: they occur in large numbers in the neighbourhood of the gabbro of the Cuillin Hills in Skye, round the eucrites of Ardnamurchan, and in Mull, Rhum and Arran. Among them are the thick sills of crinanite in South-East Arran, and sills of crinanite and olivine-dolerite in northern Skye. Included in this category also are certain radially disposed dykes of basalt and dolerite, others that are tangential to the boundaries of the major intrusions, and the remarkable inclined sheets of conical form—the **cone-sheets** of Bailey. These were described by Harker from the Cuillin Hills, where they occur as segments of cones inclined inwards towards the centre of the gabbro intrusion at angles up to 45°. It has since been discovered

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that cone-sheets are developed on an even more spectacular scale in Mull and Ardnamurchan. In the former locality they are concentrically disposed around two centres of eruption. Their injection was not restricted to one short episode as in Skye, neither are they all of basic composition: some are acid, and the intrusion of cone-sheets, which started quite early, persisted in Mull intermittently until the close of the cycle. The early cone-sheets are innumerable. They are chiefly olivine-dolerites from 30 to 40 feet in thickness, and are somewhat irregular in their attitude as they repeatedly cut one another. The majority are inclined inwards at angles of about 45° . Somewhat later in date came another series of less basic, finer-grained cone-sheets, the rocks of which include variolites, tachylites, basalts of the tholeiite type, and quartz-dolerites. These later sheets are so numerous that in a measured section the ratio of igneous rock to "screens" of country-rock is approximately 2 : 1.

Passing now to the consideration of the **regional dykes** we meet one of the most striking features of the Tertiary cycle: the dykes are remarkable not only on account of their vast numbers, but also by reason of their regularity of trend over a very extensive area. The regional direction differs little from north-west to south-east, indicating a tension acting at right angles thereto. The parallelism between this direction and the trend of Tertiary faults in other parts of Britain, and the fact that both are parallel to the extensive "deep" stretching north-westwards from the Bay of Biscay into the Atlantic, has been emphasized, and the fracturing has been explained as the result of crustal creep towards this "deep."¹ The localization of swarms of north-west dykes was no doubt due to the presence of wedges of basic magma lying at no great distance beneath the crust and causing belts of weakness. There is a marked tendency for the dykes in the neighbourhood of the plutonic centres to swing away from the regional direction and to crowd in towards the centres, while few occur in the intervening belts: the fractures occupied by the dykes crowd in towards the weak spots in the crust, rather than maintain their direction through adjacent stronger parts.² Thus the dykes are concentrated into "swarms": these comprise the Skye, Rhum, Mull, Islay, Arran, Mourne and Carlingford-Slieve Gullion swarms.

Although chiefly concentrated in the Hebrides, in the adjacent parts of the mainland of Scotland and in northern Ireland, Tertiary dykes are also found in northern England, for example, the Cleveland dyke; occasionally in the Midlands, such as the nepheline-olivine-dolerite of Butterson in North Staffordshire, while in North

¹ Evans, J. W., Presidential Address, *Quar. Journ. Geol. Soc.*, **81** (1925), especially pp. ci-ciii.

² Evans, J. W., *op. supra cit.*

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Wales they have been recorded from Anglesey¹ and various localities in Carnarvonshire.² Those of North England have been shown to converge upon the Mull swarm,³ while the Welsh dykes clearly belong to the Antrim swarm.

The dykes were formed from basic magma which, rising rapidly, suffered little contamination or differentiation. Essentially the dyke-rocks are dolerites or basalts according to their grain-size. Those

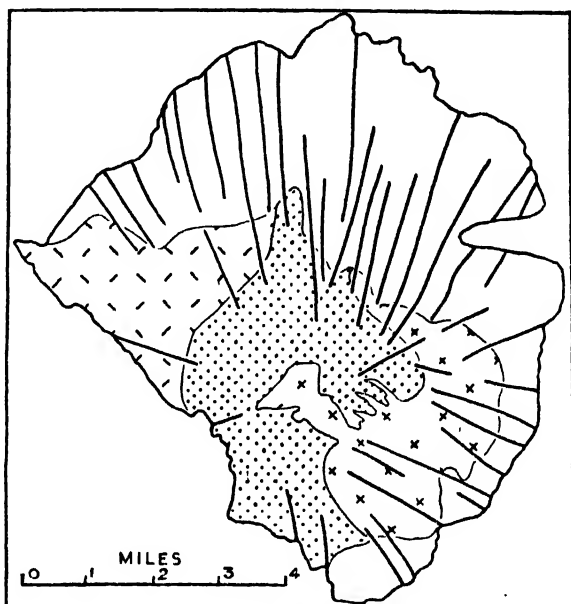


FIG. 148

Sketch-map showing relation between radial dyke-swarm and plutonic intrusions in Rhum. Ultrabasic, spotted; basic, crosses; granite, dashes; Torridonian, blank.

of the thinner dykes are indistinguishable, under the microscope, from basaltic flows, and in a representative series from this suite, all stages from basalt-glass (tachylyte) to quite coarse ophitic dolerite may be found. In addition to normal olivine-bearing and olivine-free types, others bear evidence of a tendency on the part of the magma to migrate towards the alkali pole: common augite is replaced by deep purple-brown titanaugite, while there is a moderate amount of analcite, or other zeolite, in the interstices

¹ Greenly, E., "Geology of Anglesey," *Mem. Geol. Surv.* (1919), pp. 684-90.

² Matley, C. A., *Quar. Journ. Geol. Soc.*, 69 (1913), p. 525.

³ Holmes, A., and Harwood, H. F., "The Tholciite Dykes of the North of England," *Min. Mag.*, 22 (1929), p. 1.

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between the labradorite and other earlier formed minerals. These analcite-dolerites constitute the Crinan type of Flett, and it should be noted that no means of distinguishing between the "crinanites" in the north-west dykes (presumably of Tertiary age) and identical rocks found in close association with the Permian intrusions of the Midland Valley of Scotland has yet been discovered. Among the intrusive basalts, two types may be specially mentioned. One, first described from the Cumbrae Islands between Arran and the mainland, is a porphyritic basalt with large pale-coloured phenocrysts of anorthite set in a groundmass of laths of labradorite, enstatite and augite, with an abundant glassy base.¹ The second type is an andesitic basalt with intersertal texture (tholeiitic), the type-occurrence being the tholeiite of the Brunton dyke in Northumberland. These are of essentially the same composition as the rather coarser quartz-dolerites. In contrast to the other dyke-swarms, that of Mourne consists largely of intermediate (andesitic) rocks.²

Among the latest minor intrusions are some of ultrabasic composition, which occur chiefly in radiating dykes in or near to the basic plutonic complexes. Petrographically they resemble the peridotites which were intruded at the commencement of the plutonic phase, and include picrites (with anorthite), augite-peridotites and dunites.

We have finally to notice a few scattered intrusions of rare rock-types occurring on the outskirts of the Tertiary Province. These are of more strongly sodic character than the majority of the normal rocks of Tertiary age in this country, and include the nosean-phonolite that forms the Wolf Rock off the Cornish coast, the sodic granites of the island of Rockall,³ the riebeckite-microgranite of Ailsa Craig, and the riebeckite-trachyte of Holy Isle, Arran.

It remains to mention two other complexes which may be appropriately considered here.

Lundy Island⁴ lies off the Somerset coast, well within range of the Armorican granites of south-western England. The island consists almost entirely of granite, cut by a varied assemblage of dykes. Petrologically the granite is much like the West of England types, the earlier intrusions being sodi-potassic, with perthitic orthoclase the dominant feldspar; but there is a larger proportion of albite-oligoclase in the later types. In certain points of detail the

¹ Tyrrell, G. W., *Geol. Mag.* (1917), pp. 305-15, 350-6.

² Tomkeieff, S. I., and Marshall, C. E., "The Mourne Dyke Swarm," *Quar. Journ. Geol. Soc.*, **91** (1935), p. 251. See also Tomkeieff, S. I., and Marshall, C. E., "The Killough-Ardglass Dyke Swarm," *Quar. Journ. Geol. Soc.*, **96** (1940), p. 321.

³ Sabine, P. A., "The Geology of Rockall, North Atlantic," *Bull. Geol. Surv. Gt. Br.*, **16** (1960), pp. 156 and 178.

⁴ Dollar, A. T. J., "The Lundy Complex: its Petrology and Tectonics," *Quar. Journ. Geol. Soc.*, **97** (1941), p. 39.

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granites recall the Tertiary intrusives of the Mourne Mountains, particularly in textural similarities: the occurrence of miarolitic texture, with small beryls in the cavities, among the normal minerals. Of the dykes over nine-tenths are basalts and dolerites—some are olivine-bearing, some quartz-bearing, some are classed as tholeiites, while another link with the Tertiary Province is afforded by the pitchstones and occasional tachylytes like those of the Hebrides. Less acid dykes occur, some being trachytic and others, termed "orthophyres," are microsyenitic.

Secondly, on maps showing the Tertiary dykes, a swarm unrelated to any of the visible volcanic centres crosses Islay and Jura. The inference is that a concealed complex lies a little distance from the coast of Jura.

With the close of the Eocene cycle, igneous activity ceased in Britain. Looking back over this brief account, one is struck by the variety of the rock-types occurring among the British rocks, by the comparative perfection of the record so far as some of the cycles are concerned, and on the other hand, by the many problems connected with the genesis, age and relationship of these rocks which still await solution.

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